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CHEMICAL DATA ON DESERT VARNISH

BY CELESTE G. ENGEL AND ROBERT P. SHARP

ABSTRACT

Desert varnish forms a dark coating up to 0.10 mm thick on the exposed surfaces of many stones and outcrops in southern California deserts. Wet chemical analyses were made of varnish, the underlying weathered rind, and fresh rock for a rhyolite and two andesites. The principal elements in varnish are O, H, Si, Al, Fe, and Mn, and the last two give the deposit its distinctive physical characteristics. H_2O, Fe_2O_3, and especially MnO show the greatest enrichment. Field observations and a number of partial analyses indicate that the best varnishes are on fine-grained rocks relatively rich in Fe and Mn.

Spectrographic analyses were made of 22 varnishes, 14 rocks, 8 soils, and 5 samples of air-borne material. In the varnishes Ti, Ba, and Sr are by far the most abundant trace elements, followed by Cu, Ni, Zr, Pb, V, Co, La, Y, B, Cr, Sc, and Yb. Cd, W, Ag, Nb, Sn, Ga, Mo, Be, and Zn were recorded in some but not all varnishes. The trace-element content of all varnishes is similar, and the variations recorded are related to differences in the local geology. Most trace elements are considerably enriched in varnish—Cu and Co especially, and Ni, Pb, Ba, Cr, Yb, B, Y, Sr, and V.

The chemical data suggest that (1) varnish on stones seated in soil or colluvium is derived largely from that material, (2) varnish on large bedrock exposures come from weathered parts of the rock, (3) air-borne material is probably a minor contributor.

The formation of desert varnish is primarily a weathering process involving the solution, transportation, and deposition of Mn and Fe in particular and a host of trace elements. Most of these elements are derived from local sources, and the small amount of movement required can occur by transport in solution or possibly by ionic diffusion through moisture films. Dew may be as important a source of moisture as rain. Organic agents, such as bacteria, may cause deposition of varnish, but this has not yet been demonstrated. In the desert, evaporation and the catalytic action of MnO_2 should be capable of performing the task.

The rate of varnish formation varies widely with local conditions. Hundreds and thousands of years may be required to form a dark coating in some situations, but at one locality in the Mojave Desert a good varnish formed on the surface stones of an alluvial deposit in 25 years. Although the widespread evidence of varnish deterioration may be due to climatological change, conditions in some parts of this desert area are currently favorable to varnish formation.

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Introduction and Acknowledgments

Many rocks exposed in the deserts of southwestern United States have a black to brown surface coating, commonly called desert varnish. The term desert varnish is employed in this paper because it is well established, although film, patina, or lacquer are preferred by others (Lucas, 1905; Blackwelder, 1948; 1954, p. 14). This feature has attracted attention and comment for more than 100 years (Blake, 1857, p. 230–231). The objective of this paper is to integrate laboratory and field data pertinent to the nature, source, origin, and rate of formation of this coating. Earlier ideas on desert varnish are summarized by Linck (1901, p. 330–332), Lucas (1905, p. 13–15), Hume (1925, p. 143–161), and Laudermilk (1931, p. 51–54). The data and conclusions presented in the following pages pertain only to varnish in the Colorado and Mojave deserts of southern California, although analogous or identical coatings, possibly of diverse origin, occur in other areas, situations, and environments.

William Otto, A. E. J. Engel, George Tunell, and L. T. Silver were especially generous in giving aid, advice, and criticism. Spectro-

graphic analyses and X-ray diffraction work were by A. A. Chodos and Elisabeth Godijn. Verner Schomaker and Jean Hoerni attempted an electron-diffraction identification of constituents in the varnish, and James Bonner and Sandy Siegel investigated possible organic influences. D. R. Potter made an initial spectrographic study of trace elements, and R. R. Marshall has investigated the radioactivity of varnish. The critical comments of Eliot Blackwelder, D. Foster Hewett, and C. B. Hunt are appreciated.

Description of Varnish

Physical Characteristics

Desert varnish ranges from black to various shades of brown and is shiny, greasy, or dull. Superficially, varnish on exposed rock surfaces cannot be distinguished from similar deposits along cracks, except that much surface varnish is more shiny. As Laudermilk noted (1931, p. 55), the streak is dark brown, the hardness 4.5–5, and the appearance is amorphous. The mineral composition of the varnish is unknown. X-ray photographs yield no recognizable crystalline patterns, and electron-diffraction work was unsuccessful owing to the large electrical charge that accumulated on the surface.

Field observations and some chemical data show that varnish is layered. Where the uppermost layer of black material has been removed, the underlying part is brown. Chemical analyses of successive leaches suggest a relative decrease in Mn inward from the surface.

The varnish coating tends to be most homogeneous on smooth surfaces of fine-grained rocks and less uniform on coarse-grained rocks of diverse structure and mineral composition. Phenocrysts of feldspar or quartz, veinlets of quartz or calcite, and some stones in conglomerates usually have a lighter coating than the groundmass or matrix. Layered rocks show different degrees of varnishing on different layers (Pl. 2, fig. 2). The coating tends to be denser, darker, and more shiny in hollows and indentations of the surface than on exposed edges and points.

Relations to Underlying Rock

In thin sections cut perpendicular to the rock surface, the varnish appears partly as a surface coating and partly as an impregnation (Pl. 1). Varnishlike material occurs along fractures and intergranular openings to a depth of at least 0.25 mm. Staining and impregnation of mineral grains, largely ferromagnesian and feldspar phases, extend to similar depth. As noted by Blake (1905, p. 373) and Laudermilk (1931, p. 57) a weathered rind underlies the varnish. Some of the most prominent rinds are on volcanic rocks and as much as 3 mm thick.

Varnish attains its maximum thickness of about 0.10 mm in depressions (Pl. 1, fig. 1). The average thickness of a well-developed varnish measured about 0.05 mm on an andesite and 0.03 mm on a rhyolite. A micrometric traverse on the rhyolite showed an average thickness of 0.02 mm on quartz grains and of 0.06 mm on feldspar grains. This apparent difference may be partly due to the difficulty of distinguishing between a coating and impregnation on weathered feldspar grains.

Distribution and Environmental Relations of Varnish

General Statement

Varnish is well developed and widely distributed on bedrock and on surface stones in talus, colluvial, and alluvial accumulations (Pl. 2, fig. 1) in southern California deserts. An estimated 75 per cent of the coherent bedrock exposures of this region displays a recognizable degree of varnishing, and most desert pavements have good to excellent varnish (Pl. 3, fig. 1). Entire mountain ranges tens of miles long and thousands of feet high are covered with varnish (Pl. 4, fig. 1). In such ranges the coating extends to the tops of ridges and divides, which do not receive drainage from higher slopes. Varnish is forming or has formed almost everywhere in these deserts where a source of iron and manganese is available, either in the rocks or in associated weathered debris. The topographic surface must be reasonably stable, and the rocks must be able to accept and retain the varnish.

Environmental Relations

Physical setting.—Varnish was studied in the following geological settings: (1) bedrock exposures, (2) talus accumulations, (3) colluvial mantles, and (4) alluvial deposits. No consistent variation in varnish with respect to direction of exposure was found on bedrock outcrops, but surfaces in shadow for long periods have dull varnish. Stones in talus accumulations commonly have better varnish

than the bedrock faces that supplied the debris. The surface stones of colluvial mantles may have good varnish, but at Four Buttes (Fig. 1) the varnish on adjacent talus accumulations is even denser. Stones in the colluvial mantle here are probably too frequently disturbed by movement. On alluvium the best varnish is associated with desert pavement in which the stones are closely packed and firmly seated. Surfaces frequently disturbed by floods, creep, or frost heaving do not have well-varnished stones, and rock surfaces subject to spalling, slabbing, or the separation of joint blocks are poorly varnished. Nothing was seen to indicate that the varnish of this region is formed on stones while beneath the surface of the ground (Sioma, 1938), although subsurface stones in alluvium are known to acquire encrustations of hydrous Mn and Fe oxides.

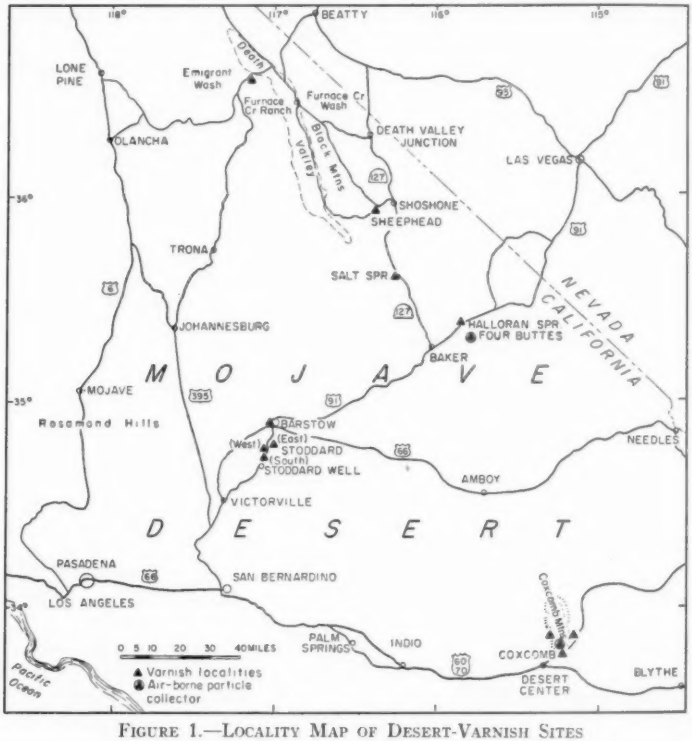

FIGURE 1.—LOCALITY MAP OF DESERT-VARNISH SITES

Varnished stones in desert pavement.—Many of the varnish specimens studied were obtained from desert pavements. Stones in such pavements are seated in the ground, and the buried part commonly has a coating ranging from dull brown to bright orange brown (Lucas, 1905, p. 8; Walther, 1924, p. 174; Laudermilk, 1931, p. 63) that contrasts vividly with the black varnish on the exposed surface (Fig. 2). The character of this bottom coating varies with lithology, nature of the soil, and probably with other less obvious factors. The minimum depth of burial required is 0.25 to 0.50 inch, but below those depths the bottom coating is uniform. Buried stones do not display a corresponding coating, but it is found on surface stones devoid of varnish, such as fragments of vein quartz.

At the ground level on embedded stones is a band of especially dark, dense, shiny varnish, usually 0.25 to 0.75 inch wide, herein referred to as the ground-line band. It extends at the most 0.25 inch above and 0.25 to 1.5 inches below ground level, depending upon the shape of the stone. A subsurface depth of 0.25 inch is common on overhanging surfaces, 0.5 inch

on near-vertical faces, and up to 1.5 inches on surfaces sloping gently outward (Fig. 3). Both the upper and lower edges of the ground-line band are gradational, but the transition

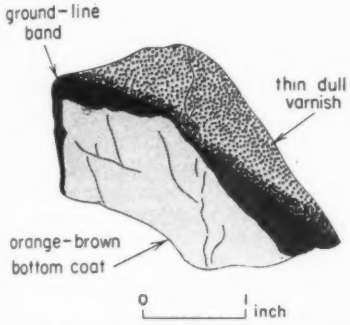

FIGURE 2.—SKETCH OF BOTTOM COAT, GROUND-LINE BAND, AND VARNISH
Specimen is from a desert pavement patch at West Stoddard locality.

materials, some tuffs, and coarse homogeneous granitic rocks subject to rapid granular disintegration. Most relatively pure limestone and dolomite are too soluble even in the desert to maintain a surface stable enough for good varnishing (Pl. 2).

The characteristics of a rock or mineral surface that permit it to accept and retain a varnish have much to do with the present distribution and preservation of this coating. Something more than stability of surface is required as shown by the general lack of varnish on fragments of vein quartz. Some cherts and flints are well varnished (Lucas, 1905, p. 8; Clements and Clements, 1953, p. 1199), probably because the weathered surface is porous. The ideal rock for development and preservation of varnish in this region (1) is relatively rich in iron and manganese or exists where these elements are plentiful, (2) is not susceptible to rapid chemical or physical weathering, and (3) has a surface with some degree of roughness

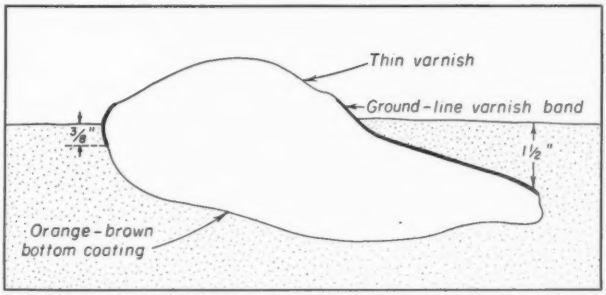

FIGURE 3.—RELATIONS OF GROUND-LINE VARNISH BAND ON PARTLY EMBEDDED STONE

occurs rapidly over 1 mm or less. The trace of these edges is irregular, reflecting in part unevenesses in the surface of the stone, and along the bottom edge dendritic outgrowths are common but not universal. Pebbles 0.5 to 1 inch in diameter, resting on the ground or only shallowly embedded therein have dark shiny varnish all over. Some fragments of vein quartz display a ground-line band.

General lithologic relations.—In the southern California deserts, good varnish has been observed on a variety of igneous, metamorphic, and sedimentary rocks including: basalt, andesite, rhyolite, diorite, gabbro, agglomerate, metatuff, metarhyolite, argillite, quartzite, phyllite, schist, gneiss, hornfels, sandstone, conglomerate, graywacke, chert, and impure carbonate rocks. Rocks that are usually not well varnished include soft friable sedimentary or porosity. Fine-grained basic lavas and some fine-grained metamorphic rocks have the best varnishes observed.

Climatological factors.—Well-varnished rock surfaces are more abundant in southern California deserts than in near-by areas that have cooler and moister climates. At one locality in this desert region varnish formed within the last 25 years, as later described, under conditions of 4 inches mean annual precipitation and a mean annual temperature between 60° and 65°F. These relations indicate that conditions favorable to varnish formation and preservation are to be found within the limits of the present southern California desert climatological environment of 1.5 to 5 inches precipitation and mean annual temperatures of 60° to 70° F.

The rate at which moisture is supplied, the

duration of the moist condition, and the temperature prevailing may be more important than total precipitation. A long gentle rainfall can be more effective than a much larger torrential downpour. Furthermore dew is not recorded in most precipitation figures, but it may play a significant role in varnish formation (Linck, 1901, p. 334–335; Hobbs, 1917, p. 31; Goldschmidt, 1954, p. 633–634). Since rock surfaces cool rapidly by outgoing radiation in the clear desert atmosphere, they constitute exceptionally favorable sites for condensation. Duvdevani (1949; 1953) shows that dew contributes significantly to moisture supplies in the deserts of Israel, and relative humidities up to 90 per cent have been recorded in California deserts even in summer (Thompson, 1929, p. 74).

In southern California deserts, good varnish exists in areas with a mean annual temperature of 60° to 70°F., July means of 80° to 100°F., and July mean maxima of 100° to 115° F. The maximum diurnal range is between 30° and 45° F. (Thompson, 1929, p. 73), and direct solar radiation can bring the temperature of varnished rock surfaces to at least 150° F. The optimum temperature conditions for varnish formation probably lie within the extremes cited. High temperature favors fixation of chemical substances in the varnish, and the temperature conditions immediately following precipitation are likely to be especially significant. Solar radiation may also be a direct factor if photochemical processes are involved in the precipitation and fixation of varnish elements.

Biological factors.—Some varnish elements may be derived from pollen, decayed plants, or other organic material (White, 1924, p. 415, 420), or varnish formation may involve the activity of living organisms such as lichens, algae, or bacteria (Laudermilk, 1931, p. 58–66; Blackwelder, 1954, p. 14; Hunt, 1954). In so far as this is true, the biological environment must influence varnish formation. Some evaluation of the role of biochemical processes is attempted in a following section of the paper.

Locality Descriptions

Introductory statement.—The specimens analyzed come from localities scattered over an area of roughly 20,000 square miles (Fig. 1), range in altitude from 540 to 3000 feet, involve many different rock types, and represent all the geological settings previously specified. Present environmental conditions at these sites do not necessarily correspond to those under which the varnishes were developed.

Halloran Spring locality (andesite).—Remnants of andesitic dikes and flows dipping gently westward form low outcrops (Pl. 3, fig. 3) on the floor of Halloran Wash 75 yards north of U. S. Highway 91 about 12.5 miles east of Baker, California. They are a quarter of a mile east of Halloran Spring in the NE ¼ of T. 15 N., R. 10 E., at elevation 3000 feet, as shown on the Old Ivanpah quadrangle (1/250,000) and the new Halloran Spring quadrangle (1/62,500). The best varnish is on float boulders lying on the southeast colluvial slope below the lava outcrops. This rock is at the base of Hewett's (1956, p. 103–105) early basalt flows of probable Quaternary age. The rock is a dark-gray, fine-grained, dense to locally scoriaceous andesite with sparsely scattered crystals of clear glassy andesine up to 3 inches long. A thin section shows pheno-

PLATE 1.—MICROPHOTOGRAPHS OF DESERT VARNISH

FIGURE 1.—Thin section cut normal to surface of varnished arkosic quartzite from Furnace Creek Wash, Death Valley. Shows thicker accumulation in depressions and irregular penetration along intergranular boundaries. Note scale line.

FIGURE 2.—Thin section cut normal to surface of varnished Stoddard rhyolite Varnish layer appears overly thick owing to photographic rendition of irregular brownish impregnation in underlying rock.

FIGURE 3.—Thin section cut normal to surface of varnished andesite from the Four Buttes locality. Varnish layer exceptionally uniform in thickness.

PLATE 2.—VARNISH IN RELATION TO LITHOLOGY

FIGURE 1.—Varnish on stones of different lithologies, surface of Emigrant Wash fan, Death Valley. Unvarnished stones are carbonate rocks.

FIGURE 2.—Impure carbonate boulder with varnish on somewhat siliceous layers. Emigrant Wash fan, Death Valley.

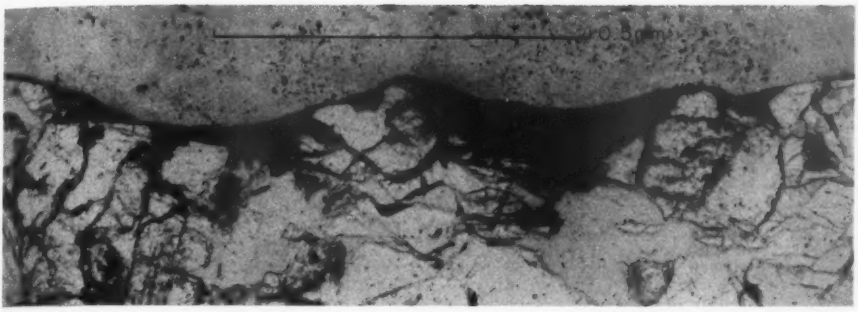

FIGURE 1

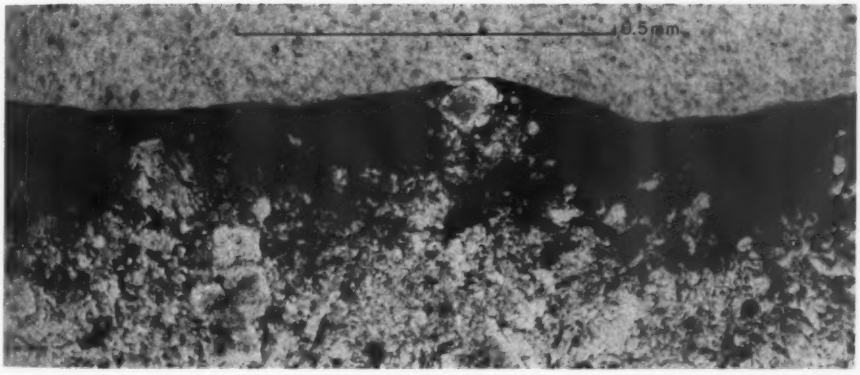

FIGURE 2

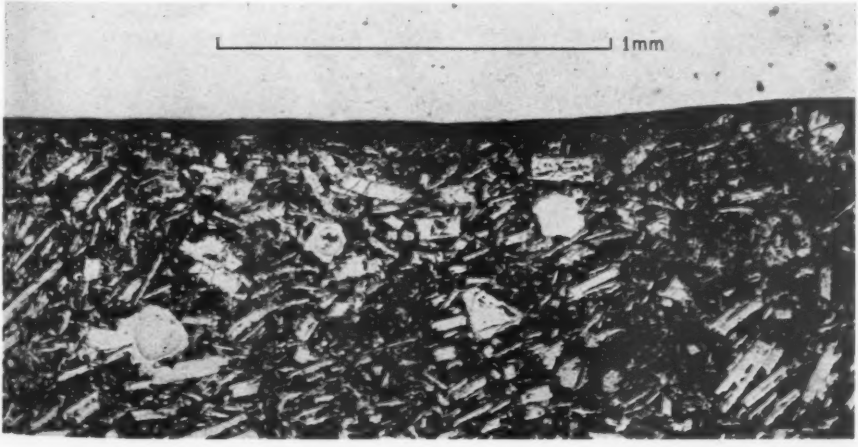

FIGURE 3

MICROPHOTOGRAPHS OF DESERT VARNISH

FIGURE 1

FIGURE 2

VARNISH IN RELATION TO LITHOLOGY

FIGURE 1

FIGURE 2

FIGURE 3
DESERT-VARNISH LOCALITIES

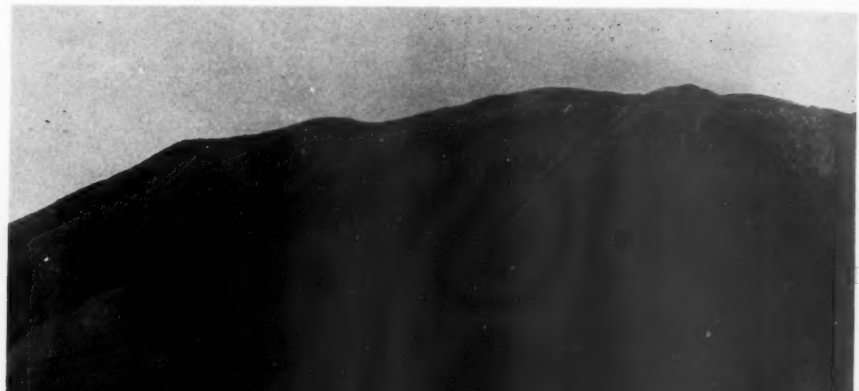

FIGURE 1

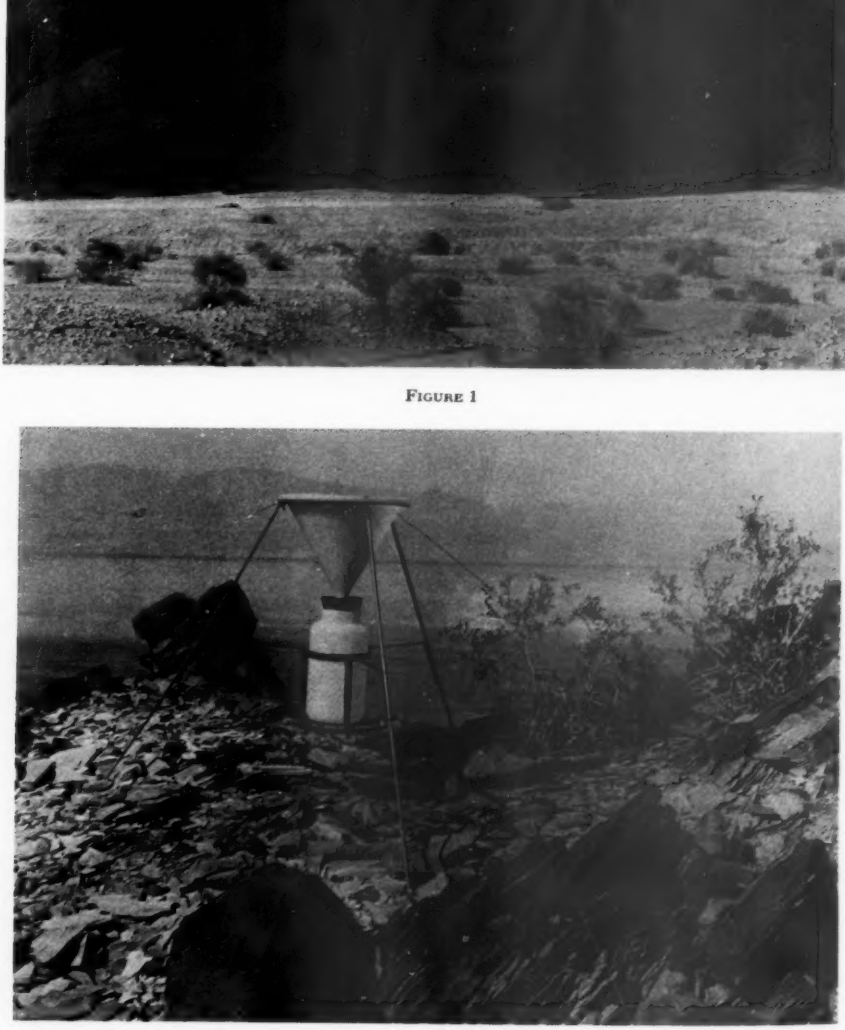

FIGURE 2

VARNISHED BEDROCK AND AIR-BORNE MATERIAL COLLECTOR

crysts of pyroxene and olivine set in a matrix of fine-grained plagioclase, pyroxene, and glass (?). In the field this rock has been identified as a basalt, but a chemical analysis (Table 1) shows high K_2O (2.07 per cent) and a somewhat low CaO (6.78 per cent), which means that it is actually an andesite. The varnish is dense, black, shiny, and relatively thick. The rind of altered rock under the varnish is light and about 0.5 mm thick.

Sheephead Mountain locality (andesite).—This site is at the north end of Sheephead Mountain in the Ibex Hills 150 yards southeast of the highway through Salsberry Pass, as shown on the Shoshone quadrangle in the NE ¼ of Sec. 2, T. 21 N., R. 5 E., at elevation 2600 feet. The specimen analyzed was part of a large boulder lying on a steep colluvial slope facing northwest. The varnished boulders come from outcrops of lava, 200 to 300 feet higher on the hillside, that are interbedded in the Funeral fanglomerate of Pliocene-Pleistocene (?) age (Noble and Wright, 1954, p. 149; Pl. 7).

The rock is a dark-gray, dense, medium-grained, slightly porphyritic andesite. A modal analysis based on 1000 counts gives a composition in volume per cent of andesine 65, pyroxene and olivine 30, and glass 5. The varnish coating is black and glossy with an underlying well-developed brownish weathered rind 2–3 mm thick.

Salt Spring locality (quartzite).—Near Salt Spring, 28 miles north of Baker, California, along State Highway 127, a small hilly outlier from the northeast part of the Avawatz Mountains is composed of well-varnished quartzites and related slightly metamorphosed rocks of sedimentary origin (Pl. 4, fig. 1). The collection site is on a steep bedrock slope facing east about 300 yards west of the highway in the NW ¼ of Sec. 32, T. 18 N., R. 7 E., at elevation 630 feet (Avawatz Pass quadrangle). The specimen analyzed comes from a bedrock outcrop of the Wood Canyon Formation of Cambrian age (B. W. Troxell, personal communication). It is a buff, medium-grained impure quartzite with distinct bedding and local cross lamination. As seen in thin section it has a granoblastic texture and is composed about 75 per cent of quartz, with some biotite and feldspar and minor amounts of colorless mica, zircon, apatite, epidote, zoisite, and pyrite. A partial chemical analysis has been made of this rock and its varnish (Table 4). The dark-chocolate-brown varnish is moderately thick, shiny, and well preserved. The underlying rock shows no megascopically discernible weathering rind.

Coxcomb Mountains localities (varied lithologies).—Specimens from several localities in the Coxcomb Mountains were studied. This area is shown on the Coxcomb Mountains quadrangle (33° 48′ N., 115° 17′ W.) about 20 miles northeast of Desert Center, California.

A specimen of well-varnished tuff, used for partial wet chemical (Table 4) and spectrographic analyses, comes from an excellent desert pavement on an alluvial fan at the south tip of the mountains. This site is about 100 yards north of Desert Center-Parker Road at elevation 540 feet, in the SE ¼ of the SE ¼ of the Army Engineers grid rectangle defined by 767–770 and 1190–1195. It is about 0.75 mile south of the mountain base at a spot where the fan slopes 1.5° southeast and is dissected by gullies 5–10 feet deep.

The tuff analyzed is a gray-green, fine-grained, well-layered rock which breaks easily into plates along a strong secondary cleavage. As seen in thin section it contains small angular fragments of quartz, orthoclase, and sodic feldspar set in a fine-grained matrix. This specimen has a dense coating of black varnish on exposed surfaces and a bright orange-brown coating on the buried part.

Samples of varnished metarhyolite and a fine tuffaceous sandstone were collected from bedrock exposures on the south slopes of the Coxcomb Mountains between elevations of 1300

PLATE 3.—DESERT-VARNISH LOCALITIES

FIGURE 1.—Varnished stones in patches of desert pavement at West Stoddard locality
FIGURE 2.—Abandoned power-line road with varnished stones, South Stoddard locality
FIGURE 3.—Varnished andesite flow at Halloran Spring

PLATE 4.—VARNISHED BEDROCK AND AIR-BORNE MATERIAL COLLECTOR

FIGURE 1.—Varnished Cambrian quartzite at Salt Spring locality
FIGURE 2.—Air-borne material collector in Coxcomb Mountains

and 1700 feet, about 1 mile north of the site described above. Samples of similar lithology were selected from desert pavement on the alluvial fan about 0.5 mile south of the mountain base. Spectrographic analyses were made of trace elements in the rock and varnish of these four specimens (Table 10).

Varnished stones of different lithologies were collected from the surface of an alluvial fan on the west flank of the Coxcomb Mountains, 5.5 miles northwest of the south tip of the mountains. The rocks include many of those described at the first site and some hybrid granitic types. This location is at an elevation of 1025 feet about 0.5 mile west from the mountain base. Trace elements in these specimens were determined by spectrographic analyses.

Emigrant Wash locality (varied lithologies).—A suite of varnished stones of different lithologies was collected from bouldery desert pavement on the surface of the Emigrant Wash alluvial fan on the west side of Death Valley (Pl. 2, fig. 1). The collection site is at an elevation of 760 feet just south of State Highway 190, 4¼ miles southest of Stovepipe Wells, in Sec. 16, T. 16 S., R. 44 E. (Emigrant Wash quadrangle). The fan surface is dissected to a depth of 3–8 feet and slopes 5° northeast. A variety of rocks is represented including several types of quartzite, graywacke, phyllite, gneiss, vein quartz, and a number of carbonate rocks with different amounts of noncarbonate material. Nearly all stones, except the relatively pure carbonates, are well varnished (Pl. 2, fig. 1), and most came from the Tucki Mountain area of the Panamint Range.

West Stoddard locality (rhyolite).—This locality, previously described by Laudermilk (1931, p. 55), is 4.5 miles north-northeast of Stoddard Well in the western Mojave Desert (Fig. 1). It lies in the SW ¼ of Sec. 23, T. 8 N., R. 2 W.[1] of the Barstow quadrangle. The varnished stones studied are in patches of desert pavement (Pl. 3, fig. 1) at an elevation of 2900 feet on a smooth alluvial fan sloping 2° northeast. Mean annual precipitation, partly snow, is about 4 inches and comes principally in winter. Mean annual temperature is between 60° and 65° F. Varnish has formed in this area within the last 25 years under these climatological conditions.

According to Bowen (1954, Pl. 1) the bedrock area shedding debris to the study locality is a coarse quartz monzonite with quartz veinlets or lenses. However, the quartz monzonite is intruded by many subparallel rhyolite, aplite, and granophyre dikes, and the bedrock area tributary to the varnish locality includes a complex of well-layered, diversified metamorphic rocks, principally a feldspathized metavolcanic series. If these materials are granitized Sidewinder volcanic rocks, they may contain considerable manganese (Bowen, 1954, p. 52). Most of the material except the dikes disintegrates granularly, and the stones in the pavement patches consist almost wholly of fine-grained rocks, particularly rhyolite, and fragments of vein quartz.

The varnished rhyolite analyzed (Table 1) is a dense, fine-grained light-gray rock with scattered phenocrysts of pink feldspar. Thin-section study shows phenocrysts of oligoclase and shredded masses of biotite set in a microcrystalline groundmass of plagioclase, orthoclase, biotite, leucoxene, epidote, and clinozoisite. Sericite is present in phenocrysts and the groundmass, which has a micrographic to sperulitic texture.

The degree of varnishing varies somewhat with the composition and size of the stone. Small stones have a more uniform, shinier and thicker coating than larger stones, which commonly have only a thin coating of dull varnish on their upper parts. Nearly all stones have a dense, dark, shiny ground-line band, and many small pebbles less than an inch in diameter are covered by a similar coating. Vein quartz is not varnished except in cracks or niches and at the ground level on some samples. Most of the rhyolitic stones have a bleached weathered rind 1–2 mm thick beneath the varnish.

The West Stoddard site has become something of a type locality for some aspects of the desert varnish problem. Laudermilk (1931, p. 58) attributes the occurrence of varnished stones in separate patches here to the influence of lichen colonies that caused or facilitated localized varnish deposition. On the basis of the following relations, the writers conclude that the patches of varnished stones are coextensive with areas of desert pavement and that the stones have developed and maintained a coating of varnish because they are relatively stable compared to stones in the interpavement areas. Lichens do not appear to have had anything to do with the patchy occurrence of varnished stones.

[1] Location erroneously given as T. 7 N. by Laudermilk.

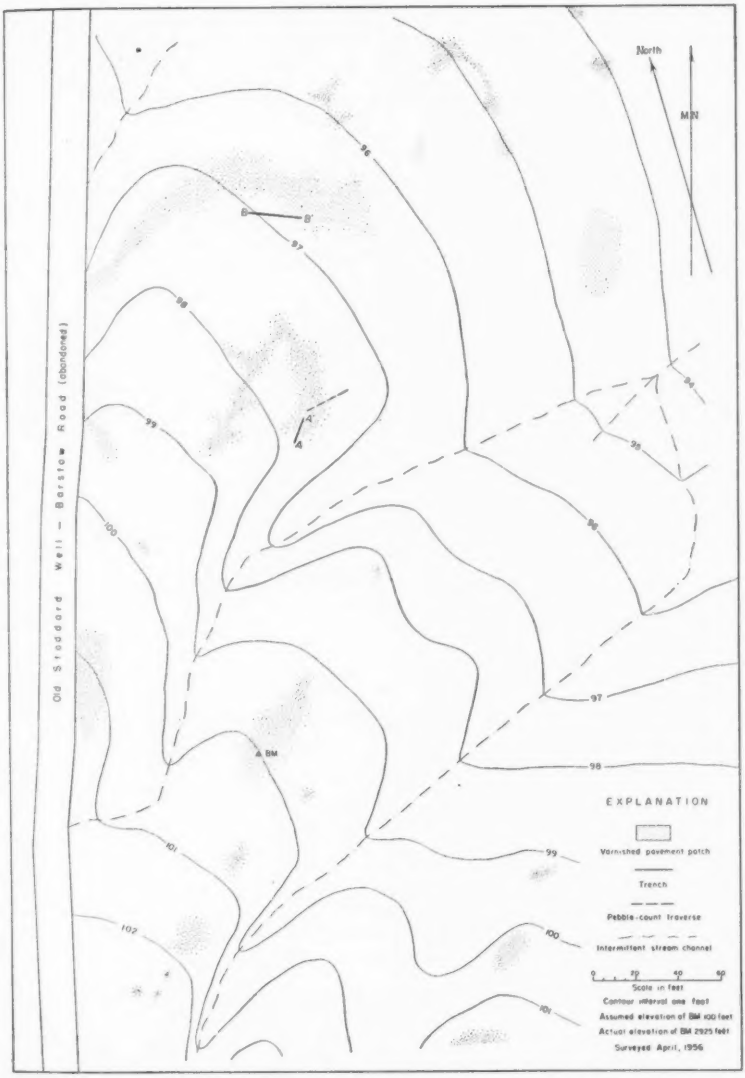

FIGURE 4.—CONTOUR MAP OF PAVEMENT PATCHES WITH VARNISHED STONES AT WEST STODDARD LOCALITY

The areas of varnished stones are relatively smooth and firm, so that walking upon them leaves no trace, and a heavy car makes a barely perceptible track. The stones are set closely together in tight mosaic style. They are flat and firmly, though shallowly, embedded in underlying fine debris. The areal ratio of stones to bare ground is 8 to 1 as established by measurements along an 11-foot traverse line. The interpavement areas of unvarnished stones consist of jumbled, bumpy ground that rises about an inch higher than the varnished patches and is so soft that a person on foot sinks in 1 or 2 inches. The areal ratio of stones to bare ground is 0.8 to 1 as measured along an 11-foot traverse. The stones in this loose ground

are tilted at all angles, are not close together, and only about 1 in 10 has varnish, which is dull and irregular in density and distribution. The size, shape, distribution, and topographic

Laudermilk (1931, p. 55) and some partial analyses by Blake (1905, p. 374), Loewe (Turner, 1909, p. 229), and White (1924, p. 418). Lucas (1905, p. 17) has made some qualitative

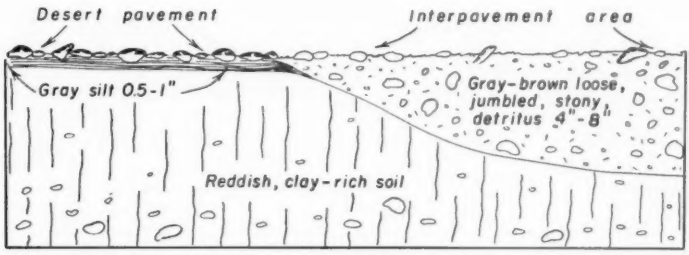

FIGURE 5.—SKETCH OF SUBSURFACE RELATIONS AT EDGE OF PAVEMENT PATCH AT WEST STODDARD LOCALITY

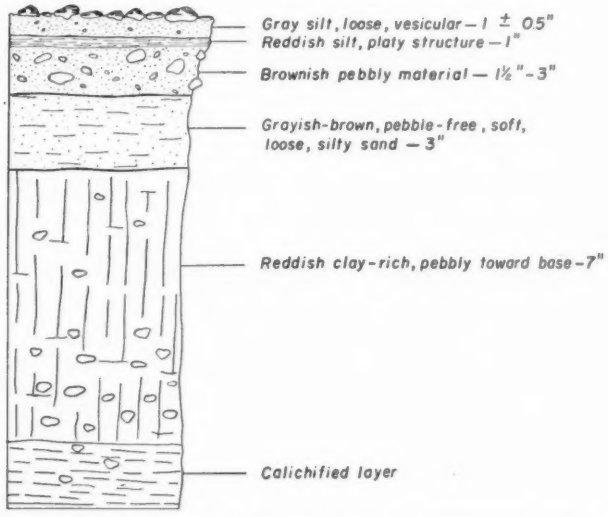
Gray silt, loose, vesicular — 1 ± 0.5"
Reddish silt, platy structure — 1"
Brownish pebbly material — 1½"-3"
Grayish-brown, pebble-free, soft, loose, silty sand — 3"
Reddish clay-rich, pebbly toward base — 7"
Calichified layer

FIGURE 6.—TYPICAL SUBSURFACE RELATIONS BENEATH PATCH OF VARNISHED STONES AT WEST STODDARD LOCALITY

relations of the areas of varnished and unvarnished stones in a representative setting are shown in Figure 4, and the subsurface relations are shown in Figures 5 and 6.

CHEMICAL ANALYSES

Introductory Statement

Data on the chemical composition of desert varnish are meager. Those found in the North American literature comprise one analysis by

determinations. In this study analyses were made by wet chemical methods of rock, rind, and varnish for the Stoddard rhyolite and the Halloran and Sheephead andesites (Table 1). Partial analyses for Fe, Mn, Ti, and P have been made of varnish-rock pairs from five other localities. Analyses were also made of two soil layers at the West Stoddard locality and of air-borne material collected in the Coxcomb Mountains. Trace elements in all these and many other samples were determined spectrographically.

TABLE 1.—CHEMICAL ANALYSES OF ROCK, RIND, AND VARNISH

(In per cent)

C. G. Engel, analyst

Stoddard rhyolite										Halloran andesite										Sheephead andesite									
Oxide weight					Atom					Oxide weight					Atom					Oxide weight					Atom				
	Rock	Rind	Varnish	Varnish*		Rock	Rind	Varnish	Varnish*		Rock	Rind	Varnish	Varnish*		Rock	Rind	Varnish	Varnish*		Rock	Rind	Varnish	Varnish*		Rock	Rind	Varnish	Varnish*
SiO_2	74.09	72.89	68.24	61.95	Si	25.22	24.66	21.10	17.74	SiO_2	50.34	46.52	39.34	28.19	Si	17.84	16.19	12.85	8.68	SiO_2	53.32	50.03	41.22	29.17	Si	18.97	17.24	13.14	8.51
TiO_2	.15	.12	.25	.33	Ti	.04	.02	.06	.07	TiO_2	2.16	2.50	2.33	2.48	Ti	.58	.65	.57	.58	TiO_2	1.54	1.71	1.98	2.42	Ti	.41	.46	.48	.53
Al_2O_3	14.61	15.76	15.33	15.93	Al	5.88	6.27	5.58	5.38	Al_2O_3	18.08	17.77	16.72	15.28	Al	7.54	7.29	6.45	5.55	Al_2O_3	17.51	18.72	19.37	21.26	Al	7.35	7.61	7.28	8.16
Fe_2O_3	.30	.87	2.52†	4.29	Fe^{+++}	.08	.22	.60	.92	Fe_2O_3	2.48	11.98	14.56†	18.56	Fe^{+++}	.65	3.15	3.58	4.31	Fe_2O_3	2.87	5.22	11.81†	15.23	Fe^{+++}	.73	1.35	2.84	3.35
FeO	.38	.25	.....	...	Fe^{++}	.10	.06	....	....	FeO	7.18	n.d.	.....	...	Fe^{++}	2.13	....	....	....	FeO	5.05	3.88	.....	...	Fe^{++}	1.50	1.12	....	...
CaO	.95	.82	.64	.31	Ca	.35	.30	.21	.09	CaO	6.78	6.34	4.78	2.76	Ca	2.58	2.38	1.68	.91	CaO	8.26	7.85	6.36	4.46	Ca	3.15	2.90	2.17	1.39
MgO	.03	.01	tr	.00	Mg	.02	.00	.00	.00	MgO	4.85	4.60	5.18	5.47	Mg	2.56	2.38	2.53	2.51	MgO	5.02	4.83	3.80	2.58	Mg	2.68	2.49	1.81	1.12
MnO	.00 .01**	.03	1.62	3.21	Mn	.00	.00	.43	.78	MnO	.18	.90	6.11	11.97	Mn	.06	.27	1.70	3.13	MnO	.09	.17	4.35	8.61	Mn	.02	.04	1.17	2.13
P_2O_5	.33	.18	.10?	—	P	.10	.06	.02	—	P_2O_5	.83	1.09	1.77	2.68	P	.25	.33	.49	.70	P_2O_5	.50	.70	.91	1.32	P	.15	.21	.25	.33
Na_2O	3.90	3.82	3.68	3.43	Na	2.58	2.49	2.20	1.90	Na_2O	4.49	3.93	2.84	1.27	Na	3.03	2.65	1.80	.76	Na_2O	3.69	3.43	2.76	1.82	Na	2.55	2.28	1.71	1.03
K_2O	5.05	4.52	3.36	1.64	K	2.19	1.96	1.32	.60	K_2O	2.07	2.02	1.70	1.31	K	.94	.90	.71	.52	K_2O	1.86	1.65	1.66	1.42	K	.81	.73	.67	.53
H_2O-	.03	.24	.56	8.85	H	.62	1.41	9.61	16.86	H_2O-	.16	.16	1.20	9.98	H	1.90	4.54	11.55	20.52	H_2O-	.06	.17	2.08	11.66	H	.86	4.09	12.73	22.73
H_2O+	.25	.38	4.12?							H_2O+	.60	1.84	4.12							H_2O+	.30	1.63	3.94						
Total	100.07	99.89	100.42	99.94						Total	100.20	99.65	100.65	99.95						Total	100.07	99.99	100.27	99.95					
Cation						37.18	37.45	41.13	44.34	Cation						40.06	40.73	43.91	48.17	Cation						39.18	40.52	44.25	49.81
0						62.82	62.55	58.87	55.66	0						59.94	59.27	56.09	51.83	0						60.82	59.48	55.75	50.18
Total						100.00	100.00	100.00	100.00	Total						100.00	100.00	100.00	100.00	Total						100.00	100.00	100.00	99.99

* Corrected for 50 per cent rock contamination in sample

† Mn interfered with FeO determination; Fe is given as Fe_2O_3

** From spectrographic analysis

Preparation of Samples for Chemical Analyses

Sampling of varnish without contamination from the underlying rock proved impossible. Varnish is easily removed in solution by a mixture of hydrogen peroxide and hydrochloric acid, but the difficulty of evaluating the contamination resulting from differential solution of constituents in the rock and other considerations related to chemical procedures made mechanical methods of sampling more desirable. A Vibro-Tool with a grinding point of hard tantalum carbide was used to abrade varnish from the surface of the specimen. Although the work was done with great care under a binocular microscope, underlying rock material was inevitably removed with the varnish. The approximate degree of contamination by rock material in a mechanically derived sample was determined by treating with a weak solution of hydrochloric acid and hydrogen peroxide until all visible varnish was removed. The residuum, consisting of mineral fragments from the rock, constituted 50 per cent by weight of the original sample. Therefore, varnish analyses in the following tables have been corrected for a 50 per cent contamination where the composition of the underlying material is known. The degree of contamination must vary with the composition, hardness, and weathered condition of the rock, thickness of varnish coating, and smoothness of the scraped surface. In addition, differential mineral contamination may result because of greater contributions by soft or weathered minerals. Preservation of varnish varies widely, and this also affects the sample if some constituents have been removed preferentially. The reader should keep in mind that the chemical analyses, although accurate within themselves, are of a contaminated material, and that this contamination tends to mask rather than emphasize the chemical peculiarities of varnish.

Altered rinds were sampled after the surface had been ground down to eliminate all traces of varnish. Rock samples were obtained well below the weathered rind, ground in an agate mortar, and sieved through silk to −100- and +150-mesh size.

Wet Chemical Analyses

Analytical procedures.—Modification of classical chemical methods was made through the use of procedures of rapid analysis formulated by Shapiro and Brannock (1952; 1956). SiO_2, Al_2O_3, CaO, and MgO were determined by classical gravimetric methods employing 1.0000-gram samples of rock and rind and .5000-gram samples of varnish. These samples were fused with Na_2CO_3 in a platinum crucible, the fusion dissolved in HCl and analyzed for the above elements. FeO was determined on a .5000-gram sample boiled gently in H_2SO_4 and HF in a large platinum crucible. The resulting solution was titrated with potassium dichromate.

Total Fe (as Fe_2O_3), MnO, TiO_2, and P_2O_5 were determined on .4000-gram samples of rock and rind and .2000-gram samples of varnish. These samples were placed in platinum crucibles and dissolved in H_2SO_4 and HF on a steam bath. The resulting solution was increased to 200 ml with distilled water. Aliquots were taken to determine the above elements on a Beckman Model B spectrophotometer. Another aliquot was used to determine Na_2O and K_2O on a Beckman Model DU photometer with flame attachment. H_2O- represents moisture driven out of the sample at 110°C., and H_2O+ was determined in a Penfield tube using a flux of lead oxide. All reagents were CPS Baker Analyzed. Standards for use in comparison on the spectrophotometer were made from materials produced and tested at the Bureau of Standards, Washington, D. C.

Principal chemical constituents of desert varnish.—In terms of atom per cent, the principal constituents of the varnishes analyzed, corrected for 50 per cent rock contamination, are O, H, Si, Al, Fe, and Mn, except for the Stoddard rhyolite in which Si is second instead of third, and Na is fifth displacing Fe and Mn to the sixth and seventh positions respectively (Table 2).

In oxide weight per cent, corrected for 50 per cent rock contamination, the principal constituents are SiO_2, Al_2O_3, H_2O, Fe_2O_3, and MnO (Table 2). In all samples SiO_2 is highest, but at Halloran Fe_2O_3 displaces Al_2O_3 from the second position, and H_2O, usually third, drops to fifth. At Stoddard, Na_2O displaces MnO from the fifth to the sixth position. MgO, CaO, Na_2O, K_2O, P_2O_5, and TiO_2 follow in different order in different samples.

Some of these differences may arise from improper correction for contamination or from selective contamination by specific minerals. The high Na_2O at Stoddard suggests that considerable Na-rich plagioclase feldspar, possibly softened by weathering, was removed in the sampling. The elements of greatest significance are those that are more abundant in

varnish than in rock. This relation can be expressed in terms of an enrichment ratio, obtained by dividing the amount of an oxide in varnish by the amount in the associated rock. Values less than unity indicate a decrease, and values greater than unity an increase, within the varnish.

TABLE 2.—ORDER OF ABUNDANCE OF ATOMS AND OXIDES IN VARNISH*
(In per cent)
C. G. Engel, analyst

Stoddard rhyolite		Halloran andesite		Sheephead andesite	
Atom	Oxide weight	Atom	Oxide weight	Atom	Oxide weight
O	SiO_2	O	SiO_2	O	SiO_2
Si	Al_2O_3	H	Fe_2O_3	H	Al_2O_3
H	H_2O	Si	Al_2O_3	Si	H_2O
Al	Fe_2O_3	Al	MnO	Al	Fe_2O_3
Na	Na_2O	Fe	H_2O	Fe	MnO
Fe	MnO	Mn	MgO	Mn	CaO
Mn	K_2O	Mg	CaO	Ca	MgO
K	TiO_2	Ca	P_2O_5	Mg	TiO_2
Ca	CaO	Na	TiO_2	Na	Na_2O
Ti	.....	P	Na_2O	K	K_2O
P	.....	Ti	K_2O	Ti	P_2O_5
Mg	.....	K	.....	P	.....

* Corrected for 50 per cent rock contamination.

From these ratios (Table 3) it appears that the most distinctive varnish element is Mn, with an enrichment ratio for MnO in the samples analyzed of 66.2 to 292. Iron is a second diagnostic varnish element, with an enrichment ratio for Fe_2O_3 of 1.7 to 6.1. Water has an enrichment ratio of 13.8 to 60 for $+H_2O$ and of 12.5 to 41.0 for $-H_2O$. Other oxides showing enrichment are TiO_2 (1.2 to 2.2) and P_2O_5 (2.0 to 3.3). The enrichment of Al_2O_3 in two samples is too slight to be significant. All other oxides are depleted in the varnish compared to the rock.

Since Mn and Fe are the most characteristic varnish elements, partial analyses were made of a number of rock-varnish pairs to determine the variation of these elements with local environment and potential sources. Analyses for Ti and P were included in some instances. In general, the data (Table 4) suggest that varnishes high in Fe and Mn formed on rocks relatively rich in these elements. Some of the exceptions may be due to differences in contamination or to deterioration of the varnish. The high Fe and Mn on Barstow rhyolite (Table 4) results from thick varnish and low rock contamination. The low enrichment ratios for the Coxcomb tuff and Coxcomb limestone probably reflect a high degree of contamination from these soft rocks. The data of

TABLE 3.—ENRICHMENT RATIOS FOR THE PRINCIPAL OXIDES IN VARNISH COMPARED TO ROCK*

	Halloran andesite	Sheep-head andesite	Stoddard rhyolite	Average ratio
MnO	66.2	86.7	292.0	148.0
H_2O^-	13.8	60.0	32.8	35.1
H_2O^+	12.5	41.0		
Fe_2O_3	2.0	1.7	6.1	3.3
P_2O_5	3.3	2.0	...	1.8
TiO_2	1.2	1.3	2.2	1.6
Al_2O_3	0.8	1.1	1.1	1.0
MgO	1.1†	0.5	...	0.8
SiO_2	0.6	0.6	0.8	0.7
Na_2O	0.3	0.5	0.9	0.6
K_2O	0.6	0.8	0.3	0.6
CaO	0.4	0.5	0.3	0.4

* Varnishes corrected for 50 per cent rock contamination.

† Possible interference by manganese.

Table 4 strengthen the impression gained from field observations that the darkest, thickest, and most lustrous varnishes occur on rocks richer in Fe and Mn, as earlier noted by Lucas (1905, p. 23).

Chemical composition of potential sources of varnish material.—The obvious sources for the substances composing varnish are the associated rocks, near-by weathered debris, and air-borne material. Chemical analyses of varnished rocks (Table 1) indicate that many of them contain the required elements, and the direct relation between the amounts of Mn and Fe in a specific varnish and its associated rock suggests that in these instances the rock contributed directly or indirectly to the varnish.

At the West Stoddard locality Mn was not detected in the rhyolite by wet chemical means, and spectrographic analysis showed only 90 ppm, so that addition of Mn and other elements from an exterior source seems likely. The same applies to fragments of Mn-free vein quartz at this locality with varnish in cracks, hollows, or in a ground-line band. The surface soil at the West Stoddard locality with 5.39 per cent Fe_2O_3 and 0.16 per cent MnO is a considerably better source for these elements than the rock

with only 0.7 per cent Fe_2O_3 and less than 0.01 per cent MnO (Table 5).

Air-borne material collected in the Coxcomb Mountains also contains all the essential elements of varnish (Table 5). Such material certainly settles onto rock surfaces, and it could contribute if some mechanism exists for incorporating the proper elements into the varnish. However, the composition of various varnishes should be more uniform and should not show a strong relation to the immediate environment if air-borne material were the sole or even the principal source of varnish substances.

In summary, soil or other disintegrated and partly decomposed surface debris in which varnished rocks are seated appear to be suitable sources for the critical elements Mn and Fe, as suggested by Blake (1905, p. 375). For large bedrock exposures and open talus accumulations devoid of fine material, where a contribution from the ground mantle of weathered debris seems impossible, the rock appears a likely source. The variation in Fe and Mn content displayed by different varnishes in direct relation to the abundance of these elements in the associated rock indicates that the rock contributes either directly or indirectly to the varnish. In so far as the major constituents are concerned, air-borne material can be a contributor, but probably not a dominant one. Whatever the source of material, the process of varnish formation produces an exceptional enrichment in Mn and H_2O and a moderate enrichment in Fe. Ti and P are less consistently enriched.

TABLE 4.—PARTIAL CHEMICAL ANALYSES OF ROCK-VARNISH PAIRS

(In per cent)

C. G. Engel, analyst

	Stoddard rhyolite†		Halloran andesite††		Sheephead andesite††		Stoddard gabbro†**	
	Rock	Varnish*	Rock	Varnish*	Rock	Varnish*	Rock	Varnish*
Fe_2O_3	.70	4.29	10.46	14.56	8.43	11.84	6.75	11.66
MnO	.00 (.01) spec.	3.21	.09	6.11	.09	4.35	.10	2.74
TiO_2	.15	.33	1.54	2.33	1.54	1.98	.77	.68
P_2O_5	.33	.....	.50	1.77	.50	.91	.13	1.72
	Barstow rhyolite††† ***		Coxcomb tuff†		Coxcomb limestone†		Salt Spring quartzite†††	
	Rock	Varnish	Rock	Varnish*	Rock	Varnish*	Rock	Varnish
Fe_2O_3	2.01	24.84	4.80	7.56	5.63	5.70	5.77	11.90
MnO	.07	8.49	.07	3.30	.06	1.52	.05	8.00
TiO_2	.....	.....	.65	.82	.79	1.86	.85	.67
P_2O_5	.....	.....	.03	.20	.26	.08	.24	.29

Environments: † Alluvial †† Colluvial ††† Bedrock

* Corrected for a 50 per cent rock contamination

** Stoddard gabbro is from a locality in the NE ¼ of Sec. 24, T. 8 N, R 2 W; 1.75 miles east-northeast of West Stoddard locality.

*** Barstow rhyolite is from a rhyolite plug near railroad station in city of Barstow.

Composition of altered rinds.—If varnish comes from the underlying rock, one might expect to find some depletion of the necessary elements within the weathered rind or within the rock beneath it. A simple calculation shows that the rind, if leached of Fe and Mn, could easily supply the amount of these elements in varnish on rocks like the Sheephead andesite. An evaluation of this possibility is provided by analyses of rock, rind, and varnish (Table 6) and by partial analyses of a series of samples from the core to outer edge of seemingly fresh rock in a 10-inch fragment of Sheephead andesite (Fig. 7; Table 7).

These data do not support the idea that the immediately underlying rock or rind supplied the varnish. In terms of atom per cent, O, Si, Ca, Mg, Na, and K progressively decrease

TABLE 5.—COMPOSITION OF SOIL, VARNISH, AND ROCK AT WEST STODDARD LOCALITY AND OF AIR-BORNE MATERIAL FROM COXCOMB MOUNTAINS

(In per cent)

C. G. Engel, analyst

	West Stoddard locality				Air-borne material, Coxcomb Mountains
	Soil: gray, Surficial, depth 1 inch	Soil: reddish, depth, 5 inches	Varnish (corrected)	Rock (rhyolite)	
SiO_2	59.49	53.63	61.95	74.09	51.00
TiO_2	.99	.67	.33	.15	.85 (spec)
Al_2O_3	17.39	18.07	15.93	14.61	21.33
Fe_2O_3	4.42	.....	.....	.30	3.18
FeO	.87	.....	.....	.38	.....
Total Fe (as Fe_2O_3)	5.39	5.86	4.29	.70	.....
CaO	3.30	2.33	.31	.95	6.76
MgO	1.93	2.43	.....	.03	2.50
MnO	.16	.11	3.21	.00	.04
				.01 (spec)	(spec)
P_2O_5	.41	.05	.....	.33	?
Na_2O	3.80	3.69	3.43	3.90	4.69
K_2O	3.22	2.79	1.64	5.05	6.71
H_2O-	1.51	4.80	8.85	.03	.06
	CO_2?	CO_2?			
H_2O+	2.80	6.18		.25	?
Total	100.29	100.61	99.84	100.07	97.12+?

TABLE 6.—COMPOSITION OF ROCK, ALTERED RIND, AND VARNISH IN ATOM PER CENT

C. G. Engel, analyst

	Stoddard rhyolite			Halloran andesite			Sheephead andesite		
	Rock	Rind	Varnish	Rock	Rind	Varnish	Rock	Rind	Varnish
Si	25.22	24.66	21.10	17.84	16.19	12.85	18.97	17.24	13.14
Ti	.04	.02	.06	.58	.65	.57	.41	.46	.48
Al	5.88	6.27	5.58	7.54	7.29	6.45	7.35	7.61	7.28
Fe^{+++}	.08	.22	.60	.65	3.15	3.58	.73	1.35	2.84
Fe^{++}	.10	.06		2.13			1.50	1.12	
Ca	.35	.30	.21	2.58	2.38	1.68	3.15	2.90	2.17
Mg	.02	.00	.00	2.56	2.38	2.53	2.68	2.49	1.81
Mn	.00	.00	.43	.06	.27	1.70	.02	.04	1.17
P	.10	.06	.02	.25	.33	.49	.15	.21	.25
Na	2.58	2.49	2.20	3.03	2.65	1.80	2.55	2.28	1.71
K	2.19	1.96	1.32	.94	.90	.71	.81	.73	.67
H	.62	1.41	9.61	1.90	4.54	11.55	.86	4.09	12.73
O	62.82	62.55	58.87	59.94	59.27	56.09	60.82	59.48	55.75
Total	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00

from rock through rind to varnish, and Fe, Mn, H, and usually Ti and P progressively increase. Aluminum increases relatively in the rind and decreases in the varnish. These figures are for varnish samples uncorrected for rock contamination, and the trends cited are actually greater than appear in Table 6. Since the rind is richer than the rock in the critical elements, Fe and Mn, it looks as though varnish material were added to the rind rather than taken from it, or that these elements were deposited there on the way outward from the rock. This seems unlikely since analyses of the fresh rock do not show any leaching gradient measurable by wet chemical analyses (Table 7). Varnish on the specimens analyzed did not come from the rock beneath, although it may have come from some closely associated part of the same rock body.

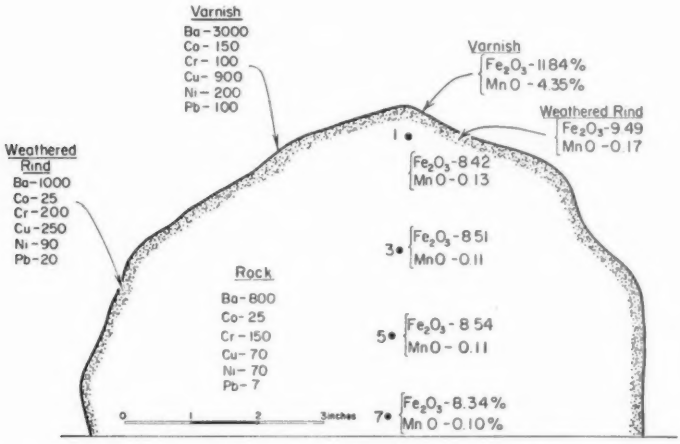

FIGURE 7.—VARIATIONS OF SELECTED ELEMENTS IN ROCK, RIND, AND VARNISH ON SHEEPHEAD ANDESITE FRAGMENT

Details shown in Tables 7 and 13. Trace elements in parts per million

TABLE 7.—PARTIAL CHEMICAL ANALYSES OF SAMPLES FROM CORE TO EDGE OF SHEEPHEAD ANDESITE FRAGMENT, OXIDE WEIGHT PER CENT

C. G. Engel, analyst

	Sec. 7	Sec. 5	Sec. 3	Sec. 1	Rind	Varnish
	core- - - - - - - - - - outward					
TiO_2	1.40	1.40	1.43	1.43	1.71	1.98
Total Fe	8.34	8.54	8.51	8.42	9.49	11.84
MnO	0.10	0.11	0.11	0.13	0.17	4.35
P_2O_5	0.46	0.48	0.51	0.51	0.70	0.91

General Summary of Wet Chemical Data

(1) Wet chemical analyses show that the most characteristic elements of varnish are Mn and Fe.

(2) MnO has an enrichment ratio of 66 to 292, H_2O of 13 to 60, and Fe_2O_3 of 2 to 6, in varnish, compared to the rock on which it occurs.

(3) Soil, rock, or air-borne material contain the essential elements of varnish. Where it can be a contributor, soil appears to be a more likely source of Fe and Mn than rock. Where soil cannot contribute, rock is a better source than air-borne material, but in the specimens analyzed the varnish must have come from some other part of the rock body than that immediately underlying the varnish.

Trace-Element Relations

Introduction.—Spectrographic analyses of trace elements were made with the hope of defining sources of the varnish material. The 22 varnishes analyzed came from a variety of rocks including andesite, rhyolite, tuffaceous sandstone, tuff, metarhyolite, several quartzites and graywackes, phyllite, gneiss, impure carbonate rocks, and hybrid granitic types. Various environmental settings and most of

the localities shown on Figure 1 are represented. In addition, 14 rocks, 8 soils, and 5 air-borne specimens were analyzed.

Varnish samples were obtained by abrasion and are contaminated with rock material for which no correction has been made; any element showing enrichment in varnish is present in amounts greater than reported. Sensitivities of the spectrographic analyses are listed in Table 8. The large amount of iron in varnish interferes with the determination of Ga, In, and Mo. Reproducibility on the same spectrographic plate was usually within a few per cent of the amount of an element present, but a few differences as great as 30 per cent were reported. Reproducibility on different spectrographic plates is less reliable, and maximum differences reported for some elements range between 30 and 100 per cent. For these reasons the spectrographic data are used primarily to indicate general trends and broad relations, and more reliance is placed upon group behavior than on individual variations.

TABLE 8.—SENSITIVITIES OBTAINABLE IN ROUTINE SPECTROGRAPHIC ANALYSES OF ROCKS

Element	ppm	Element	ppm	Element	ppm	Element	ppm
Ag	0.5	Cr	10	Mo	2	Ta	200
As	200	Cs	50*	Nb	20	Th	200
Au	20	Cu	1	Ni	1	Ti	1
B	10	Ga	1	Pb	10	Tl	50
Ba	1	Ge	20	Pt	50	U	500
Be	0.8	In	10	Rb	10*	V	1
Bi	10	K	2000 (5)*	Sb	200	W	100
Ca	1	La	100	Sc	5	Y	10
Cd	20	Li	30 (1)*	Sn	5	Yb	1
Co	1	Mn	1	Sr	1	Zn	100
						Zr	2

* Requires additional exposure.

Average trace-element content of varnish.—Elements such as Mn, Fe, Si, Al, Ca, Mg, Na, K, and P, usually present in amounts greater than 1 per cent, are not treated here as they are more accurately determined by wet chemical methods. Fifteen trace elements were recorded in all varnishes, and 9 additional elements were found in some of the samples analyzed. The elements occurring in all varnishes are subdivided into three groups on the basis of abundance. Ti (3900),[2] Ba (2200), and Sr (700) are by far the most abundant; next come 9 elements ranging from Cu (260) to B (60), followed by 3 elements of less than 50 ppm average abundance (Table 9). Of the elements recorded in some but not all varnishes, most are present in less than 100 ppm.

TABLE 9.—AVERAGE TRACE-ELEMENT CONTENT OF 22 VARNISH SAMPLES

A. A. Chodos and E. Godijn, analysts

			Ppm	
			Mean	Range*
Elements present in all varnishes	Group I >300 ppm	Ti	3900	1000–9000
		Ba	2200	500–4000
		Sr	700	100–2500
	Group II 50–300 ppm	Cu	260	30–900
		Ni	210	40–1500
		Zr	200	90–350
		Pb	180	90–300
		V	130	40–420
		Co	110	30–300
		La	110	40–200
		Y	80	50–200
		B	60	30–170
	Group III <50 ppm	Cr	40	10–140
		Sc	20	10–50
		Yb	6	3–15
Elements present in only some varnishes		Cd	..	0–300
		W	..	0–600
		Ag	..	0–80
		Nb	..	0–70
		Sn	..	0–30
		Ga	..	0–15
		Mo	..	0–10
		Be	..	0–5
		Zn	..	0–200

* Maximum range given is that recorded in different samples on a single spectrographic plate.

Among the elements of moderate abundance the amount of Zr (200) is about the same as in the associated rocks, and Cu (260) and Ni (210) although high are erratic in range. Among the elements not present in all varnishes, W is possibly a contaminant, and most others are present in small amounts; caution must be used in attaching much significance to them. The amount of Zn in varnish samples cannot be determined, because its lines are obscured by interference from other elements,

[2] Figures in parentheses represent the average content in parts per million in 22 varnishes.

TABLE 10.—TRACE-ELEMENT CONTENT OF ROCK-VARNISH PAIRS*

A. A. Chodos and E. Godijn, analysts

	Stoddard rhyolite (fan)		Halloran andesite (colluvial)		Sheephead andesite (colluvial)		Coxcomb tuff (fan)		Coxcomb tuffaceous sandstone (bedrock)		Coxcomb tuffaceous sandstone (fan)		Coxcomb metarhyolite (bedrock)		Coxcomb metarhyolite (fan)	
	Rock	Varnish	Rock	Varnish	Rock	Varnish	Rock	Varnish	Rock	Varnish	Rock	Varnish	Rock	Varnish	Rock	Varnish
Ag	—†	1	—	—	—	—	—	—	—	—	—	—	—	—	—	—
B	—	10	10	50	35	35	35	40	20	25	10	25	p. Tr	15	10	10
Ba	400	2000	300	~5000	800	~3000	1000	2000	1200	~600	2400	3400	2800	4300	2100	3200
Be	Tr	—	Tr	6	—	Tr	Tr	—	—	—	—	—	—	—	—	—
Co	—	90	20	200	25	150	5	70	4	90	7	130	15	90	20	80
Cr	2	75	65	65	150	100	15	20	60	10	25	35	30	35	80	140
Cu	3	30	15	250	70	900	2	70	25	150	15	180	70	850	120	760
Ga	5	6	15	10	10	15	8	10	6	—	5	3	7	6	7	10
La	—	85	50	60	55	100	—	100	—	120	—	360	120	170	120	170
Mo	—	5	—	8	—	4	—	3	—	<5	—	6	—	Tr	<5	—
Nb	20	15	10	15	—	10	—	—	—	—	—	—	—	—	—	—
Ni	—	40	40	150	70	200	7	50	7	40	6	370	15	1500	30	940
Pb	10	100	pos T	150	7	100	45	100	—	140	30	220	40	160	50	140
Sc	3	5	20	30	30	30	20	20	15	10	9	15	25	20	25	30
Sn	?	9	—	30	—	20	—	8	p. Tr	—	—	15	—	15	—	15
Sr	60	250	1000	1500	1500	1500	200	400	310	120	870	1000	330	200	460	400
Ti	500	1000	10,000	9000	8000	9000	2500	2500	2400	1400	2100	2600	3900	3300	4200	5400
V	15	45	90	150	100	200	85	90	100	35	80	140	290	160	450	420
Y	30	50	55	200	30	80	50	100	30	70	35	120	55	70	45	60
Yb	2	3	4	15	3	7	4	7	1	2	1	4	2	2	1	2
Zr	100	90	200	200	100	200	150	150	200	100	200	270	350	220	250	260

* All rock-varnish pairs are on the same spectrographic plate. The first four pairs are on one plate and the last four pairs on another. Values in parts per million.

† Dashes indicate elements looked for but not found; As, Au, Bi, Cd, Ge, In, Pt, Sb, Ta, Th, Tl, U, W also looked for but not found.

Among the substances that stand relatively high on the list of trace elements in varnish the chemical coherence of Ba with Mn is well known, and the behavior of Sr is similar. Goldberg (1954, p. 252–257) shows that iron oxides are accompanied by Ti, Co, and Zr, and manganese dioxides by Cu and Ni in ocean-bottom deposits. He attributes this to the scavenger action of these compounds.

Although the trace-element content of individual varnishes is different, there is a broad similarity in the samples analyzed. This resemblance prevails irrespective of lithology, location, or details of the local environmental setting. In all varnishes Ti, Ba, and Sr are most abundant, and the trace elements of moderate and minor abundance are essentially the same, although the order of abundance may be different. These relations may be due in part to the fact that most of the abundant minor elements are chemical cogeners in a deposit of this composition, and the scavenging action of hydrated iron and manganese oxides may be only one of several mechanisms by which these concentrations are produced.

Trace-element contents of rocks compared to varnishes.—Spectrographic analyses of rock and varnish are compared for a rhyolite, two andesites, two tuffaceous sandstones, two metarhyolites, and a tuff, from four separate localities (Table 10). A strong resemblance but not an exact correspondence exists between the trace elements in rock and varnish. All elements recorded in the rock are found in the varnish in greater abundance. Some varnishes contain elements not recorded in the rock, but most of these are present only in a few parts per million, except for La, Co, and Ni which may approach 100 ppm. All these elements may be present in such small quantities in the rock that they escape detection, or they may have been contributed from some outside source.

Certain trace elements display a wide range of abundance in different varnishes (Table 9), and this appears to depend on the composition of the varnished rock (Tables 10 and 11). For example, the varnish on a cobble of white quartzite from the Emigrant Wash fan in Death Valley (Table 11) is exceptionally low in Sr, Cr, V, and Ba and moderately low in

TABLE 11.—VARIATIONS OF TRACE-ELEMENT CONTENT OF VARNISHES ON DIFFERENT ROCKS*
A. A. Chodos and E. Godijn, analysts

Halloran andesite (colluvial)		Sheephead andesite (colluvial)		Stoddard rhyolite (fan)		Emigrant Wash white quartzite (fan)		Coxcomb tuffaceous sandstone (bedrock)		Coxcomb metarhyolite (bedrock)	
Ti	9000	Ti	9000	Ba	2000	Ti	1500	Ti	1400	Ba	4300
Ba	~5000	Ba	3000	Ti	1000	Ba	1000	Ba	~600	Ti	3300
Sr	1500	Sr	1500	Sr	250	W	500	Zn	190	Ni	1500
Cu	250	Cu	900	Pb	100	Zr	350	Cu	150	Cu	850
Co	200	Ni	200	Co	90	Pb	150	Pb	140	Zr	220
Y	200	V	200	Zr	90	Sr	100	La	120	Sr	200
Ni	150	Zr	200	La	85	Co	100	Sr	120	La	170
Pb	150	Co	150	Cr	75	Cu	80	Zr	100	Pb	160
V	150	Cr	100	Y	50	Y	80	Co	90	V	160
Cr	65	La	100	V	45	B	70	Y	70	Co	90
La	60	Pb	100	Ni	40	La	70	Ni	40	Y	70
B	50	Y	80	Cu	30	Ni	50	V	35	Cr	35
Sc	30	B	35	Nb	15	Cd	40	B	25	Sc	20
Nb	15	Sc	30	B	10	V	30	Cr	10	Sn	15
Yb	15	Sn	20	Sn	9	Nb	15	Sc	10	B	15
Ga	10	Ga	15	Ga	9	Cr	10	Mo	<5	Ga	6
Mo	8	Nb	10	Sc	5	Sc	6	Yb	2	Yb	2
Be	6	Yb	7	Mo	5	Mo	5			Mo	Tr.
		Mo	4	Yb	3	Ag	4				
		Be	Tr.	Ag	<1	Be	3				

* Data for the Halloran, Sheephead, and Stoddard localities from the same spectrographic plate; Emigrant Wash and Coxcomb data are both from different plates. Values in parts per million.

Ni, Sc, and Ti. Varnish on the Halloran andesite is very high in Ba, Sr, Ti, Co, Y, and Yb. Varnish on the Sheephead andesite is rich in Ti, Sr, Cu, Cr and high in Ni, Co, Ba, and La. In other instances, rocks low in Cu, Co, Ni, V, and Y have varnishes notably low in these elements; the relation is especially marked for Cu (Table 10). Rocks low in Sr have varnishes relatively low in Sr, even though the enrichment ratio for this element is especially high in these instances. Ti and Zr are essentially the same in rock and varnish. In general, rocks relatively high in certain trace elements tend to have varnishes rich in those elements and in amounts beyond those that can be attributed to contamination of the varnish sample by rock material. Pb is strongly enriched in many varnishes, but lead-rich varnishes do not necessarily occur on lead-rich rocks. Cu and Co consistently show relatively high enrichment in varnish compared to rock, and Ni, Pb, Ba, Cr, B, Yb, Y, and Sr follow in about that order (Table 12).

TABLE 12.—ENRICHMENT RATIOS FOR TRACE ELEMENTS IN 11 ROCK-VARNISH PAIRS

Element	Enrichment ratio	Element	Enrichment ratio	Element	Enrichment ratio
Cu	4–35	Yb	1–7	Ti	0.6–2
Co	6–23 (∞)	B	1–6 (∞)	Sc	0.8–2.1 (∞)
Ni	3–115 (∞)	Y	1.4–3.6 (∞)	Nb	1–1.5 (∞)
Pb	2–14 (∞)	Sr	0.4–4	Ga	1–1.5
Ba	0.5–20	V	0.3–5	Zr	0.5–1.3
Cr	0.2–38	La	0.6–2 (∞)	...	.....

In brief, the strong resemblance between trace elements in varnish and in the rock upon which it occurs suggests that the rock contributes directly to the varnish or to an intermediate material, such as soil, from which the varnish is derived. The correspondence between trace elements in rock and varnish is greater for bedrock outcrops and colluvial settings than for stones on an alluvial surface. Varnish formation produces moderate enrichment in a number of trace elements and a high enrichment in a few.

Variation of trace-element content from rock through altered rind to varnish.—If varnish comes principally from the underlying rock, some depletion of trace elements should be found in the rock or in the weathered rind. Analyses to test this were made for the Stoddard rhyolite and the Sheephead andesite and for four samples of fresh rock taken progressively outward from the center of a 10-inch fragment of the andesite (Table 13).

In both specimens the trace elements Ba and Cu show a progressive and marked increase from rock through rind to varnish; the same is true for Cr, Sr, Ti, V, and Y in the Stoddard specimen and for Ni and Pb in the Sheephead. For Stoddard, B, Co, La, Mo, and Ni are not recorded in the rock but are present in rind and varnish, and in Sheephead only Sn has this relation. Nb is higher in rind than in either rock or varnish for Stoddard, but in Sheephead Cr is higher. For Stoddard, Pb shows no increase in the rind compared to the rock although it is higher in the varnish, and in Sheephead Co and Zr behave this way. Elements present in the varnish but not in rock or rind are Sn in Stoddard and Be, Mo, and Nb in Sheephead, all in very small amounts. Within the limits of analytical error, no element is less abundant in the rind than in the rock.

From the above, it seems that Co, La, Mo, Ni, and Sn could have been added to the rind from the varnish. Elements greatly enriched in the varnish compared to the rock are also enriched in the rind but generally to a smaller degree. The succession of samples from within the Sheephead andesite provides no evidence of a leaching gradient (Fig. 7). Suggestions of such a relation for B, Ba, and Cu are not strong enough to be significant. The above data give little support to the idea that trace elements in the varnish have been drawn from the underlying rock by a leaching process; rather, elements were probably added to the rind from some exterior source. The number of trace elements added to the rind is much greater at the Stoddard locality, an alluvial environment, than at the Sheephead site, a colluvial occurrence.

Trace elements in soils.—Soil and other weathered mantles are a possible source of varnish on stones resting on or partly embedded in such materials. The relatively greater amount of manganese in the soil than in the varnished rock at the West Stoddard locality suggests that the varnish was derived from the soil. It is of interest to see if the trace-element content of soil and of varnish on stones embedded therein confirms this suggestion.

Data are listed in Table 14, and the trace-

element content of the varnished rock is provided for comparison. With some exceptions, the principal trace elements, Ti, Ba, and Sr, are more abundant in the soil than in the rock. However, a number of elements found in varnishes at some localities are not recorded in the associated soils. At Emigrant Wash these include Ag, Be, Cd, Ga, La, Mo, Nb, and Pb, at Stoddard they are Ga, La, Mo, Nb, Pb, and Sn, and at Coxcomb they are Ga, La, Mo, Pb, and Sn. Except for La and Pb (70–150 ppm) these elements are present in only a few ppm in the varnish. Most of them are recorded in the varnished rock; exceptions are La, Mo, and Sn. Soil could probably furnish many but possibly not all the trace elements in varnishes unless the elements have escaped detection in the soils because of low concentrations.

TABLE 13.—TRACE ELEMENTS IN ROCK, WEATHERED RIND, AND VARNISH*

A. A. Chodos and E. Godijn, analysts

	Stoddard rhyolite			Sheephead andesite					
				Core ---------- outer edge					
	Rock	Rind	Varnish	Sec. 7	Sec. 5	Sec. 3	Sec. 1	Rind	Varnish
Ag	—†	—	<1	—	—	—	—	—	—
B	—	10	10	35	35	30	30	30	35
Ba	400	600	2000	800	800	900	900	1000	~3000
Be	T	T	—	—	—	—	—	—	T
Co	—	40	90	25	25	20	25	25	150
Cr	2	6	75	150	150	100	150	200	100
Cu	3	15	30	70	70	60	60	250	900
Ga	5	5	6	10	9	10	10	15	15
La	—	50	85	55	70	50	65	65	100
Mn	90	550	5000	400	500	450	450	550	1%
Mo	—	— or T	5	—	—	—	—	—	4
Nb	20	50	15	—	—	T	—	—	10
Ni	—	15	40	70	80	60	80	90	200
Pb	10	10	100	7	7	7	8	20	100
Sc	3	5	5	30	30	30	35	35	30
Sn	?	?	9	—	—	—	—	2	20
Sr	60	90	250	1500	1500	1500	1500	1500	1500
Ti	500	700	1000	8000	8000	8000	8000	9000	9000
V	15	20	45	100	150	150	200	200	200
Y	30	35	50	30	40	30	45	43	80
Yb	2	2	3	3	3	3	3	3	7
Zr	100	100	90	100	150	90	100	100	200

*All data from a single spectrographic plate. Values in parts per million.

†Dashes indicate elements looked for but not found, including As, Au, Ge, In, Pt, Sb, Ta, Th, Tl, U and W which are not listed

Spectrographic analysis of air-borne material.—The possibility that desert varnish consists principally if not wholly of material that settles out of the atmosphere onto exposed rock surfaces must be considered. As an approach to evaluating this possibility, the trace-element content of air-borne material was investigated.

A collector for air-borne material was established on the crest of the Coxcomb Mountains, 20 miles northeast of Desert Center (Fig. 1), near localities from which specimens were obtained for varnish analyses. These mountains are composed of a thick section of slightly metamorphosed, bedded clastic rocks of Precambrian or Paleozoic age, and for the most part well varnished. The collector was set up in a small saddle on a sharp ridge near the south tip of the mountains at an elevation of 1750 feet, 1100 feet above the alluvial fans at the mountain base. The site is in the midst of exposures of calcareous argillite, tuff, graywacke, and grits, and vegetation on adja-

cent slopes is exceedingly sparse. The apparatus consisted of a 12-inch plastic funnel leading into a rubber-stoppered plastic jar (Pl. 4, fig. 2). The funnel rim was 30 inches above ground level. Material that settled onto the funnel slid or was washed into the bottle by dew or rain. The apparatus was set up on May 28, 1955; the first sample was taken on May 13, 1956, a second on December 29, 1956, and a third on January 19, 1958.

Similar apparatus was established 110 miles to the north on top of a small lava-capped butte at the Four Buttes locality near Halloran Spring on October 6, 1956. The elevation of this site is 3645 feet, the butte rises 250 feet above the surrounding country, adjacent rocks are andesitic lavas, and the slopes below as well as the immediate surroundings are mantled with granitic debris. Vegetation consists chiefly of scattered creosote bushes. The first sample was taken here on April 5, 1957, and the second on January 18, 1958.

TABLE 14.—TRACE ELEMENTS IN VARNISH, ROCK, AND SOIL*

A. A. Chodos and E. Godijn, analysts

	Stoddard Well			Coxcomb Mountains			Halloran Spring			Emigrant Wash 760 feet elevation	
	Rock	Varnish	Soil	Rock	Varnish	Soil	Rock	Varnish	Soil	Mean of eight varnishes	Soil
Ag	—†	<1	—	—	—	—	—	—	—	16	—
B	—	10	35	35	40	35	10	50	26	83	70
Ba	400	2000	1200	1000	2000	400	300	~5000	1150	965	1100
Be	T	—	—	T	—	—	T	6	—	2	—
Cd	—	—	—	—	—	—	—	—	—	89	—
Co	—	90	10	5	70	15	20	200	10	66	10
Cr	2	75	90	15	20	60	65	65	50	41	65
Cu	3	30	40	2	70	30	15	250	30	195	50
Ga	5	6	—	8	10	—	15	10	—	5	—
La	—	85	—	—	100	—	50	60	—	70	—
Mo	—	5	—	—	3	—	—	8	—	2	—
Nb	20	15	—	—	—	T?	10	15	—	23	—
Ni	—	40	25	7	50	20	40	150	20	105	25
Pb	10	100	—	45	100	—	pos T	150	—	160	—
Sc	3	5	20	20	20	15	20	30	10	22	15
Sn	?	9	—	—	8	—	—	30	—	?	—
Sr	60	250	320	200	400	270	1000	1500	390	345	380
Ti	500	1000	6900?	2500	2500	4700	170	9000	3500	3500	4900
V	15	45	160	85	90	120	90	150	110	120	100
W	—	—	—	—	—	—	—	—	—	125	—
Y	30	50	75	50	100	59	55	200	25	70	45
Yb	2	3	3	4	7	3	4	15	<2	6	<2
Zn	—	T?	—	—	—	Interf.	—	—	—	Interf.	Interf.
Zr	100	90	1150	150	150	620	200	200	300	240	420

* Data are from three spectrographic plates. Values in parts per million.

† Dashes indicate elements looked for but not found; As, Au, Bi, Ge, In, Pt, Ta, Th, Tl, and U also looked for but not found.

In all except one instance the bottles held considerable water which had a pH of 5 to 5.5. It contained a thick growth of green to brown scum, presumably produced by air-borne algae or other micro-organisms. On the bottom were small amounts of silt and some fine sand. Moths and ants had also managed to get into the collectors. The whole mess had a fairly strong unpleasant organic odor.

The average air-borne material proved to have a trace-element content similar to the average of 22 varnish samples (Table 15). Ti, Ba, and Sr are present in corresponding order but in somewhat lower amounts in the

air-borne material. Some differences in the arrangement and grouping of other elements are apparent. In the air-borne material, Ni and Co are exceptionally low, Cu and Y are low, and Cr is high compared to average varnish. However, in so far as trace elements are concerned it would be possible to get everything found in the average varnish from air-borne material except La, Cd, Nb, Sn, and Be, of which only La is present in all varnishes. Fractionation and concentration would be required to yield the right amounts and ratios, but this is true to some degree for all possible sources.

A comparison of average air-borne material at the Coxcomb collector with the average of five varnish specimens (Table 15) from stones on the surface of an alluvial fan at the west base of the Coxcomb Mountains, 4 miles northwest of the collector site, shows a lower degree of correspondence. The air-borne debris is much lower in Co and considerably lower in Pb, Ba, Sr, Ni, and Y. It is somewhat higher in Cr and Zr but has no La, Cd, Nb, Sn, or Be, all of which are recorded in the varnish.

Similar relations exist at the Four Buttes locality where average air-borne material (Table 15) contains much less Ti and Co, considerably less Ba, Sr, Y, and Yb, and somewhat less Cu and Ni than near-by varnish at

TABLE 15.—TRACE ELEMENTS IN AIR-BORNE MATERIAL, AVERAGE VARNISH, AND LOCAL VARNISH*
A. A. Chodos and E. Godijn, analysts

Element	Coxcomb Mountains					Four Buttes				Average of all air-borne samples	Average varnish
	Air-borne material				Local varnish Average of 5 specimens	Air-borne material			Local varnish (Halloran)		
	5/13/56	12/29/56	1/19/58	Average†		4/3/57	1/18/58	Average			
Ti	2500**	2500	2600	2500	2000	2800	2600	2700	9000	2600	3900
Ba	1350	1300	1600	1400	3100	1700	1600	1650	~5000	1525	2200
Sr	400	375	420	400	1400	365	375	370	1500	385	700
Cu	160	59	78	100	180	135	77	105	250	100	260
Ni	19	17	36	25	80	88	28	60	150	40	210
Zr	165	185	230	200	140	158	155	155	200	175	200
Pb	83	57	45	60	290	275	82	180	150	120	180
V	120	104	150	125	135	225	140	180	150	150	130
Co	8	8	11	10	170	16	11	13	200	10	110
La	—††	—	—	—	130	—	—	—	60	—	110
Y	25	29	34	30	95	27	23	25	200	25	80
B	93	43	87	75	45	165	58	110	50	90	60
Cr	115	98	115	110	35	117	73	95	65	100	40
Sc	15	14	16	15	25	15	14	15	30	15	20
Yb	3	3	3	3	9	3	3	3	15	3	6
Cd	—	—	—	—	70	—	—	—	—	—	0–300
Ag	<1	<1	3	1	2	9	1	5	—	3	0–80
Nb	—	—	—	—	9	—	—	—	15	—	0–70
Sn	—	—	—	—	20	180(?)	—	—(?)	30	—	0–30
Ga	31	28	25	28	9	39	24	30	10	30	0–15
Mo	<10	—	<10	<10	6	11	<10	<10	8	<10	0–10
Be	—	—	—	—	3	—	—	—	6	—	0–5
Zn	134	57	—	100	140	202	107	155	Interf.	125	0–200

* All air-borne samples on the same spectrographic plate, varnish samples on different spectrographic plates.

† All average figures rounded off.

** All values in parts per million.

†† Dashes indicate elements looked for but not found; As, Au, Bi, Ge, Pt, and Sb were also looked for but not found.

Halloran Spring. The air-borne material contains a little more B and Ga and also Ag which is not recorded in the varnish, but it lacks La, Nb, Sn, and Be all of which are in the varnish. In trace-element content air-borne material shows a closer correspondence to average varnish than to local varnish.

TABLE 16.—TRACE-ELEMENT CONTENT OF VARNISH AS A FUNCTION OF LOCATION*

A. A. Chodos and E. Godijn, analysts

	Average of eight varnish samples on stones on Emigrant Wash fan at 760 feet elevation		Average of five varnish samples on stones on Coxcomb Mountains fan	
Group I >300 ppm	Ti	3500	Ba	3100
	Ba	1000	Ti	2000
	Sr	350	Sr	1400
Group II 50–300 ppm	Zr	240	Pb	290
	Cu	195	Cu	180
	Pb	160	Co	170
	W	125 (contamination?)	Zr	140
	V	120	Zn	140
	Ni	105	V	135
	Cd	90	La	130
	B	85	Y	95
	La	70	Ni	80
	Y	70	Cd	70
	Co	65		
Group III <50 ppm	Cr	40	B	45
	Nb	25	Cr	35
	Sc	20	Sc	25
	Ag	15	Sn	20
	Yb	6	Ga	9
	Ga	5	Nb	9
	Mo	2	Yb	9
	Be	2	Mo	6
	Sn	?	Be	3
	Zn	interference	Ag	2

* All data from a single spectrographic plate.

Influence of location on trace-element content of varnish.—To evaluate the possible influence of location on variation in trace-element content, the average of eight analyses from Emigrant Wash in Death Valley is compared with the average of five analyses from the Coxcomb Mountains, 225 miles to the south. In both localities the varnish samples are from stones of several different lithologies on the surface of an alluvial fan. Although the data in Table 16 show a general similarity of trace elements in these varnishes, distinct local differences are apparent. The general grouping of elements is comparable, but the order of arrangement within these groups is different. Ba is the most abundant trace element at the Coxcomb locality, and Ti predominates at Emigrant Wash. The relative position of Co is very different, and there are other less notable examples. These data suggest that the local geological environment influences the trace-element composition of the varnish. The differences are too great to be solely the result of rock contamination in the sample.

Summary.—The trace-element content of all varnishes is similar in broad aspects regardless of location, lithology, or details of the local environmental setting. The elements can be arranged into three groups on a basis of abundance, and with minor exceptions they tend to remain in these groups from sample to sample although their position within the group may change. Ti, Ba, and Sr are the most abundant trace elements in all varnishes. The average amounts and relative rank of 21 other trace elements in 22 varnishes are shown in Table 9.

The trace-element contents of three possible sources—rock, soil, and air-borne material—resemble the average composition of varnish, but no single source appears to contain all the elements in a varnish. Rock generally contains more of the elements than other sources, but spectrographic analyses of rock, altered rind, and varnish fail to show any leaching gradient within the rock or to lend much support to the idea that the trace elements have come directly from the immediately underlying rock. Many of the trace elements in varnish on stones in alluvial environments have probably been supplied from the adjacent soil.

Variations in the trace elements of varnish are attributed to the influence of local geological environments to which rock and soil must be major contributors. Large variations of specific trace elements in varnish reflect similar differences in the associated rocks. Air-borne material undoubtedly settles onto the surface of exposed rocks, and in so far as this material can be incorporated into the varnish it must contribute. This is supported by the fact that Mo is reported in small amounts in a number of varnishes, but among the possible sources it is recorded only in air-borne material. However, a greater uniformity in the trace-element content of varnish would be

expected if air-borne material were the sole or even a principal source.

Origin of Desert Varnish

Introduction

Theories on the origin of desert varnish basically involve secretion or accretion. In secretion, the varnish substance comes primarily from the rock upon which it rests; in accretion it has been supplied from some other source. In either instance migration is required, and organic processes may have been involved. Lucas (1905, p. 24), Blake (1905, p. 373–374), Turner (1909, p. 230), Merrill (1921, p. 245), Walther (1924, p. 176), Klute and Krasser (1940, p. 22), Twenhofel (1932, p. 78), and Kiersch (1950, p. 936) support secretion; Laudermilk (1931, p. 65–66) supports organic secretion; possibly Blackwelder (1954, p. 14) and certainly White (1924, p. 415, 420) support organic accretion. The writers believe that the varnish is primarily an accretion in the sense that it did not come from the imediately underlying rock.

Source of Materials Composing Varnish

Chemical data presented in preceding sections suggest that the principal elements, Fe and Mn, are most readily available in associated weathered debris or in the varnished rock. Trace elements in the varnish can best be supplied by the rock, but many could have come from weathered debris. Air-borne material may have supplied some of the major components and some of the trace elements, but it does not appear to be the sole source of all varnish material. Varnish on stones partly embedded in a weathered mantle has probably been derived largely from that mantle, and varnish on large bedrock exposures has come principally from weathered parts of the bedrock but not necessarily from the rock immediately underlying the varnish.

Behavior of Manganese in Varnish Formation

Since Mn is the most characteristic and most enriched constituent, knowledge of its source and behavior is basic to an understanding of varnish formation. The Mn must come initially from a bedrock source, although it may have passed through several intermediate stages before reaching the varnish. Mn is not overly abundant in earth materials, and the average igneous rock contains something less than 0.1 per cent by weight (Goldschmidt, 1954, p. 621). In igneous and metamorphic rocks, Mn is divalent and occurs primarily as a minor constituent within the structure of ferromagnesian silicates, rather than as distinct Mn mineral species. Much of the Mn in sedimentary rocks and weathered debris has been oxidized to the trivalent or quadrivalent state.

The ionic radius of Mn^{2+} (0.91Å) is larger than other divalent elements of the iron family (Goldschmidt, 1954, p. 634), and its low ionic potential (Mason, 1952, p. 137–140) makes it readily susceptible to leaching even in weak acid solutions. This is demonstrated by the rapid loss of Mn in early stages of rock weathering (Clarke, 1924, p. 492). Mn in higher valence states is more stable and must generally be reduced before going into solution. Manganous ions are more easily soluble in carbonate-bearing waters than ferrous ions, and the manganous bicarbonate, $Mn(HCO_3)_2$, is more stable in solution than the ferrous bicarbonate. Manganese is readily soluble in sulfate-bearing water, but $MnSO_4$ is a minor constituent of weathering solutions (Rankama and Sahama, 1950 p. 647). The relatively high content of P_2O_5 in varnish suggests that some of the Mn may be transported as a soluble phosphate (Lucas, 1905, p. 13–14; Walther, 1924, p. 175). Once in solution the Mn can be transported by moving films of moisture, or it can move by ionic diffusion through such films.

Mechanisms for depositing Mn onto a rock surface are less clearly understood. Deposition can be caused by increased alkalinity and an abundance of free oxygen, by organic agents, or by evaporation. Deposition from the soluble bicarbonate occurs in the presence of free oxygen in solutions of pH 7.5 or higher. This may involve conversion of the bicarbonate to $Mn(OH)_2$ and subsequent oxidation to $Mn(OH)_4$. This compound can stay in solution as a colloidal hydrosol, or it can be deposited as $Mn(OH)_4$ or MnO_2 (Rankama & Sahama 1950, p. 647). Precipitation is facilitated through removal of CO_2 by various means including organisms. MnO_2 has a strong catalytic effect on the deposition of Mn (Zapffe, 1931, p. 816, 824), and this may be an important factor in enrichment of Mn in desert varnish. Linck (1901, p. 333) suggests that ammonium nitrate and sodium chloride in the weathering solutions should speed up oxidation of Fe and Mn.

The formation of Mn-rich deposits from Fe-Mn solutions is a problem recently treated in

detail by Krauskopf (1957). The Mn/Fe ratio in average igneous rocks is about 1/60 (Goldschmidt, 1954, p. 621). Krauskopf's (1957, p. 69–70) experiments and work at California Institute of Technology (L. T. Silver, personal communication) show that Mn and Fe are artificially leached from igneous rocks in about this ratio. If natural weathering solutions are of similar behavior, the problem is to produce from such Fe-rich solutions a varnish with a Mn/Fe ratio in the neighborhood of 1/1.4.

This might come about because of: (1) exceptional deposition of Mn owing to the strong catalytic effect of MnO_2, to adsorption of Mn by colloidal iron, or to some unknown inorganic chemical influence; (2) precipitation of Mn by organisms such as bacteria; (3) deposition, either biochemical or inorganic, from solutions already enriched in Mn because of earlier precipitation of Fe; (4) leaching of Fe from the varnish subsequent to deposition. The possibility that natural weathering solutions in arid regions are initially rich in Mn should also be considered.

Deposition of Mn from Fe-bearing solutions without first precipitating most of the Fe is difficult without the aid of some special agent, such as bacteria, or some special influence, such as a catalyst. If Mn deposition occurs because of such influences, there must be some means of removing the residual Fe-enriched solution from the surface of the rock. In desert areas rapid evaporation makes this especially difficult. Mn/Fe ratios in some carbonate-bearing waters are said to be as high as 5/1 (Rankama and Sahama, 1950, p. 647), probably because of earlier removal of Fe (Krauskopf, 1957, p. 69, 73). This supports the possibility that the solutions lost much of their Fe before arriving at the site of varnish deposition. Such an explanation is more satisfactory for stones seated in a weathered surface mantle than for bedrock. In the former, Fe could be deposited in the soil (Table 5), but on bedrock it is difficult to find a place in which to dispose of the Fe. If by some peculiarity of the environment or the nature of the varnish deposit, quadrivalent Mn becomes less soluble than trivalent Fe in hydrated oxides, Mn enrichment of varnish might be produced by leaching of Fe subsequent to deposition.

From the above considerations, varnishing by solutions rich in Mn is favored over leaching of Fe from varnish after deposition. Since nothing is known about the chemical nature of the weathering solutions, the values of the pH and Eh attending deposition (Krauskopf, 1957), and the complex chemistry involved, the above is purely speculative.

Possible Role of Organisms in Varnish Formation

The possibility that organisms play a direct or indirect part in the formation of desert varnish has long been regarded with favor (White, 1924, p. 420; Laudermilk, 1931, p. 65) and remains popular (Blackwelder, 1954, p. 14; Hunt, 1954). White (1924, p. 415), impressed by clouds of air-borne pollen in southwestern Arizona, suggested that the pollen adhere to sun-heated rock surfaces and that subsequent decomposition disposes of the organic material leaving Mn and Fe as an oxide coating on the rock. Pollen are not particularly abundant in many areas of good varnish; they have not been seen on rock surfaces; and they are not known to be rich in Mn and Fe.

Laudermilk (1931, p. 65–66) suggested that lichens take up Mn and Fe from the rocks on which they grow and redeposit them on the surface as hydrated oxides of Mn and Fe. After the plant dies organic acids derived from its decomposition peptize the hydroxides allowing them to spread freely over the rock surface where they are subsequently fixed as insoluble oxides by heating and oxidation. Laudermilk (1931, p. 58) attributed the patches of varnished stones at the West Stoddard locality to the work of lichen colonies, but they are actually patches of desert pavement that need have no relation to lichens. Lichens have been observed growing on varnished rock in some localities, but contrary to Laudermilk's (1931, p. 65–66) statement, the surface beneath the lichens in these instances is well varnished. The plants seemingly took up residence on the rock after the varnish was formed. Lichens also grow on unvarnished rocks.

The six minor elements (Fe, Zn, Mn, Cu, B, and Mo) essential to plant life (Bonner and Galston, 1952, p. 49) are found in desert varnish, and in some instances considerably enriched. This is especially true for Cu and to a smaller degree for B. Mo is in many varnishes and not in the associated rock or soil, but the amount is small, less than 10 ppm. Analyses by Segal (Personal communication) showed the presence of benzene rings and lignin in a varnish sample. These facts suggest the possibility that trace elements taken up by plants have been added to the varnish. Perhaps the easiest way of doing this would be through the medium of air-borne debris which included partly decomposed plant substance. The air-

borne material collected in the Coxcomb Mountains contains all the essential minor elements. Plants may have contributed trace elements indirectly to the varnish by this means, but varnish also contains large amounts of many elements not known to be favored by plants.

Certain bacteria promote the deposition of Mn and Fe (Zapffe, 1931, p. 802, 805–813), and they may play a part in the fixation of these elements in varnish. However, it has not yet been demonstrated that the proper species of bacteria live or have lived on rock surfaces in the desert—Lucas (1905, p. 43–44), for one, is skeptical—or that biochemical precipitation of Mn has occurred or is required. Krauskopf (1957, p. 71) is explicit on this point as follows:

"That bacteria can serve as a general solution to the manganese problem, however, is hardly likely. To suppose that the manganese-precipitating bacteria were present in all the various environments where manganese deposits are found, and to suppose further that they were present to the exclusion of iron bacteria, is the sort of *ad hoc* hypothesis that cannot be completely eliminated but that seems too improbable for serious consideration."

Bacteria might have a function in causing the initial deposition of Mn from weathering solutions. Once some MnO_2 is present, its strong catalytic effect should accelerate inorganic precipitation of Mn. The possible role of bacteria in varnish formation deserves investigation, but it is only one facet of the desert-varnish problem.

Ground-Line Band and Bottom Coating

The ground-line band shows that conditions near the ground level are especially favorable to the formation of varnish. This may be due principally to the large and readily available supply of Mn in the surrounding soil. Below the level of the ground-line band, the bottom coating on the stones consists of hydrated oxides of Fe without Mn (Table 17). Conditions must become rapidly unfavorable to Mn deposition at depths in excess of about half an inch. Some of the factors possibly affecting this relation are moisture, temperature, pH, oxygen, radiation, and biological processes.

The fact that the Mn-rich ground-line band extends to greater depths on buried stone faces sloping gently outward (Fig. 4) suggests that temperature and radiation may play a role in Mn deposition. The penetrating radiation incident upon such a surface is greater than upon near-vertical or overhanging faces, both of which are varnished to respectively shallower depths beneath the ground surface. Such incident radiation could raise the temperature of the rock surface, or perhaps it produces a photochemical reaction fixing Mn. Higher temperature and greater radiation might also promote biological processes causing Mn deposition. There is no apparent reason why oxygen should be more plentiful or pH more favorable over the gently sloping rock face, but the supply of moisture might be somewhat greater.

TABLE 17.—PARTIAL CHEMICAL ANALYSES OF BOTTOM COATINGS ON VARNISHED STONES*

C. G. Engel, analyst

	Stoddard rhyolite			Coxcomb tuff		
	Rock	Varnish	Bottom coating	Rock	Varnish	Bottom coating
Fe_2O_3	0.70	2.52	2.60	4.80	6.18	7.58
MnO	0.00	1.62	tr.	0.07	1.68	0.07
TiO_2	..	..	..	0.65	0.73	0.78
P_2O_5	..	..	..	0.03	0.11	0.19

* All analyses of varnish and bottom coating uncorrected for rock contamination. Figures in oxide weight per cent.

These relations suggest that singly or together temperature, radiation, biological processes, and possibly moisture may have a part in varnish formation.

Moisture

In the desert most rains are of short duration, and the rock surfaces are wet for relatively brief periods. On one occasion the exposed surfaces of varnished stones in pavement patches at the West Stoddard locality were dry within 1 hour after a heavy winter downpour, even though the sky was still overcast and the temperature was 40°–45°F. Summer rains are characteristically episodic, and the drying must be even more rapid then. Extensive film migration or ionic diffusion probably could not occur during such short intervals, and moisture from wet ground does not climb high onto the stones once their exposed surfaces become dry.

For these reasons dew should be seriously considered as a possible factor in varnish de-

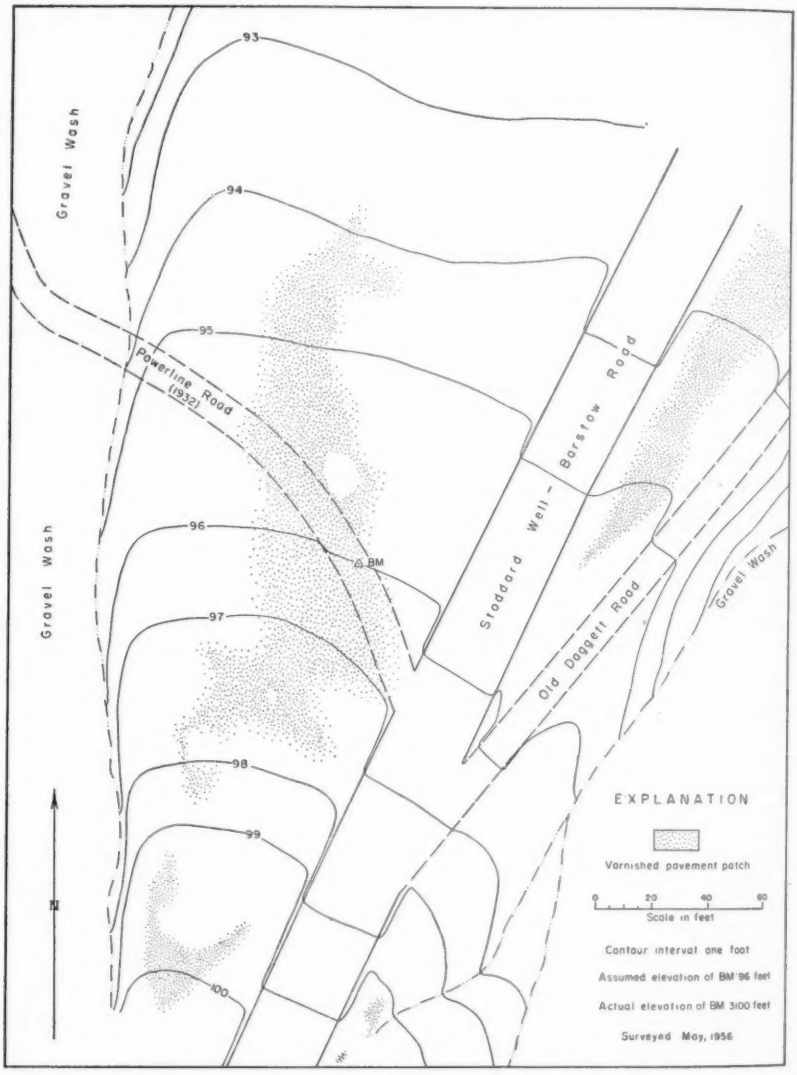

FIGURE 8.—CONTOUR MAP OF POWER-LINE ROAD, PAVEMENT PATCH, AND VARNISHED STONES AT SOUTH STODDARD LOCALITY

velopment (Linck, 1901, p. 334–335; Lucas, 1905, p. 24; Hobbs, 1918, p. 96; Goldschmidt, 1954, p. 633–634), although Hume (1925, p. 150–151) is skeptical. Even though a film of dew will evaporate rapidly early in the day, it may exist on exposed rock surfaces far longer than moisture films supplied by rain. It may be that some rock surfaces are better varnished than others because they receive more dew (Linck, 1901, p. 334).

Conclusion

The formation of desert varnish is primarily a weathering phenomenon which involves the solution of certain elements, particularly Mn

and Fe, their transportation to the varnished surface, and deposition there as a relatively stable coating. The source of the varnish is principally local, and the distance of migration need not be great. Most of the varnish substances come from chemical decomposition of the associated rocks or the surrounding mantle of weathered debris. Contribution from airborne material is probably minor. Mn and Fe may be transported in several chemical compounds. Transportation can occur by migration of solutions or by ionic diffusion through films of moisture supplied, perhaps, more by dew than by rain. Initial deposition of at least Mn may be caused by biochemical (bacterial) action, but subsequently inorganic deposition may become more important owing to the catalytic effect of MnO_2. The relatively high ratio of Mn to Fe in varnish may be due to (1) weathering solutions initially rich in Mn, (2) deposition of Fe before the solutions reach the site of varnishing, or (3) leaching of Fe subsequent to deposition of the varnish.

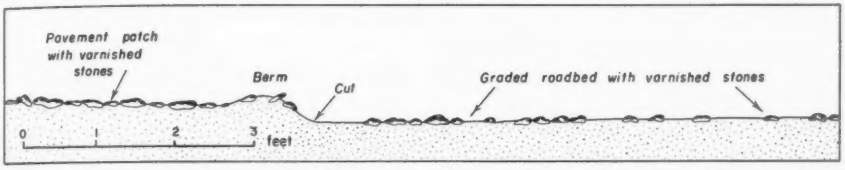

FIGURE 9.—CROSS-SECTION SKETCH OF GRADED POWER-LINE ROAD WITH VARNISHED STONES AT SOUTH STODDARD LOCALITY

TIME FACTOR IN VARNISH FORMATION

A distinction must be made between the age of a varnish and its rate of formation, for antiquity does not necessarily mean a slow rate of development. Blackwelder (1948; 1954, p. 14) cites evidence from Egypt suggesting scarcely perceptible varnish formation in 2000 years, slight (light brown) in 5000 years, and full development in 20,000 to 50,000 years. Some archeologists regard a heavy varnish on artifacts as evidence of antiquity (Skarland and Giddings, 1948, p. 119; Clements and Clements, 1953, p. 1199), but this is not a universal attitude (Movius, 1942, p. 107; Péwé, 1954, p. 55). Hunt (1954) feels that some varnish may be as old as late Wisconsin. A common tendency is to regard varnishing as a slow process and to look upon heavily varnished surfaces as relatively old. Exceptions are Setzler (1952, p. 402), who states that varnish may form in 500 years, and Klute and Krasser (1940, p. 22) who report varnish on the surface stones of Alpine moraines that must have formed within the last 40 to 80 years. Although there are many localities in the southeastern California desert in which only minor varnish development has occurred in hundreds or perhaps thousands of years, one site was discovered in which a good varnish has formed in much shorter time.

At the South Stoddard locality (Fig. 1) an abandoned road branches to the northwest off the Stoddard Well-Barstow road, opposite the junction of the old Daggett road as shown near the center of section 34, T. 8 N., R. 2 W. (Barstow quadrangle, Ed. 1934). The general setting here is an alluvial surface sloping gently northeastward with a broad wash and a rocky hill slope to the west and a corresponding hill slope to the east. Details of the local topography are shown in Figure 8. In the first 75 feet beyond the junction the abandoned road crosses an elongate patch of desert pavement with well-varnished stones (Pl. 3, fig. 2). The road was made by a blade or drag that cut at least 2 inches below the level of the desert pavement on the upslope (south side) and shoved up a low berm about 2 inches higher (Fig. 9). In the bed of the graded roadway, firmly seated in the ground, are well-varnished stones with a good ground-line varnish band and an orange-brown bottom coating. The road was made by the Southern Sierra Power Company (now California Electric Power Company) in 1931. The stones in the roadbed must have been disturbed by this operation, and it is clear that they have developed a good varnish coating with respect to their present position on the ground in the past 25 years. The conditions for varnish formation here may be ideal, since the stones offer a good surface, there is abundant iron and manganese in the adjacent soil, and the topographic situation is relatively stable. The climatic environment of about 4 inches precipitation and 60°–65° mean annual temperature must be suitable for varnish development.

The rate of varnish formation in southern California deserts is strongly affected by environmental factors such as lithology, local setting, geochemical relations, and climate. In extremely favorable situations varnish may form in less than 25 years. In others it takes hundreds and even thousands of years. The existence of a heavy varnish coating is not necessarily an indication of great antiquity, and varnish is not a good measure of time unless the various local factors pertinent to its rate of formation and preservation can be evaluated. The condition of a varnished surface as it appears at present is likely to be as much a matter of preservation as of formation, and the varnish on a specific stone may have passed through several cycles of development and deterioration (Walther, 1924, p. 178).

Deterioration of Varnish

Many small isolated patches of black shiny varnish on some rock surfaces in southeastern California deserts suggest (1) that these rocks once had a more continuous coating of similar material which has been partly destroyed, or (2) that varnish formation is controlled by highly localized conditions. The evidence indicates that both situations may occur.

Dendritic outgrowths along the margins of isolated spots of black shiny varnish indicate that they are in the process of development. Various stages in the expansion and coalescence of such spots leading to the formation of a continuous varnish coating have been observed. Since many of these patches lie in hollows, pits, or cracks, Blackwelder (1954, p. 14) suggests that they are a local development related to the greater moisture held in such places. This explanation cannot apply in all instances, for patches are also found on vertical or overhanging faces where the depressions could not harbor moisture. An alternative is that the hollows are better varnished because they are underlain by softer, more porous material that affords a better surface for varnishing. Notably different are the varnish patches of irregular outline and sharp edges (Fig. 10) which rise abruptly like scabs above the surrounding unvarnished rock surface. These patches are characterized by a lack of dendritic outgrowths and are not centered upon or even coextensive with hollows. They are the remnants of a deteriorated varnish coating.

Varnish can be worn away by abrasion of wind-blown sand, but in the examples studied independent evidence of sandblasting is lacking. Varnish can be removed by disintegration of the underlying rock as in the Rosamond Hills and in the Granite Mountains 9 miles east of Victorville. Varnish on stones of a glacial moraine formed in 1820 has been almost destroyed by rock disintegration (Beschel, 1957).

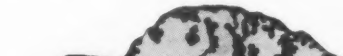

Figure 10.—Sketch of Varnish Patches on Rhyolitic Bedrock

Patches are interpreted as remnants of a deteriorated varnish coating.

The angular outline and sharp edges of the small varnish scabs described above (Fig. 10) suggest destruction by small-scale spalling or flaking. This could be the result of volume increase through progressive weathering in the altered ring beneath the varnish. In places the varnish coating is broken by circular blisters from which flaking is spreading progressively outward. Flaking away of only the outermost black layer of varnish on the Halloran andesite may also be due to alteration and expansion of the underlying rock. The destruction of varnish by rock disintegration or flaking need not indicate a change in environmental conditions, as it can be the result of a normal progression of weathering.

Removal of varnish by solution is more difficult to evaluate, although it seems to have happened in at least one instance. The upper parts of many stones on alluvial surfaces have thin dull varnishes that grade downward to a black shiny coating at the ground level (Fig. 2). Formerly these stones may have had a more extensive covering of dark shiny varnish which has been partly destroyed by solution because of some change in environmental conditions. Possibly the present dull varnish is the best that can be developed under existing conditions, if it is the product of a balance between the processes of formation and of destruction. This viewpoint is consistent with relations at the South Stoddard locality where varnish formed in 25 years is as good as that

on stones in near-by pavement patches which have been exposed to the varnishing process for a much longer time. Dull thin varnishes are not necessarily deteriorated varnishes.

Environmental changes, especially of climatic nature, must influence the rate of formation and degree of development of varnish. Hunt (1954) suggests that pluvial periods provide optimum conditions for the formation of varnish in arid regions. He recognizes varnish deterioration and by inference relates it to drier periods. Walther (1924, p. 173, 178) by contrast regards dryness rather than moisture as more favorable to varnish formation. At least four different degrees of varnish development can be recognized on the surface stones of closely associated alluvial or colluvial deposits of different ages and no great antiquity in southern California deserts. It seems unlikely that these varnishes indicate four separate climatic fluctuations. Rather, it appears that varnish formation has gone on more or less continuously within the limits of recent climatological variations, none of which may have been truly pluvial, and the degree of varnish development reflects the age of the alluvial or colluvial deposit.

In summary, some varnishes have been partly destroyed by disintegration or flaking of the underlying rock material resulting from a normal progression of weathering. Other varnishes, with features possibly attributable to destruction by solution, may indicate environmental change, but it has not been demonstrated that they are truly deteriorated varnishes. Instead, they may be the product of a steady-state balance between processes of varnish formation and varnish destruction under present conditions.

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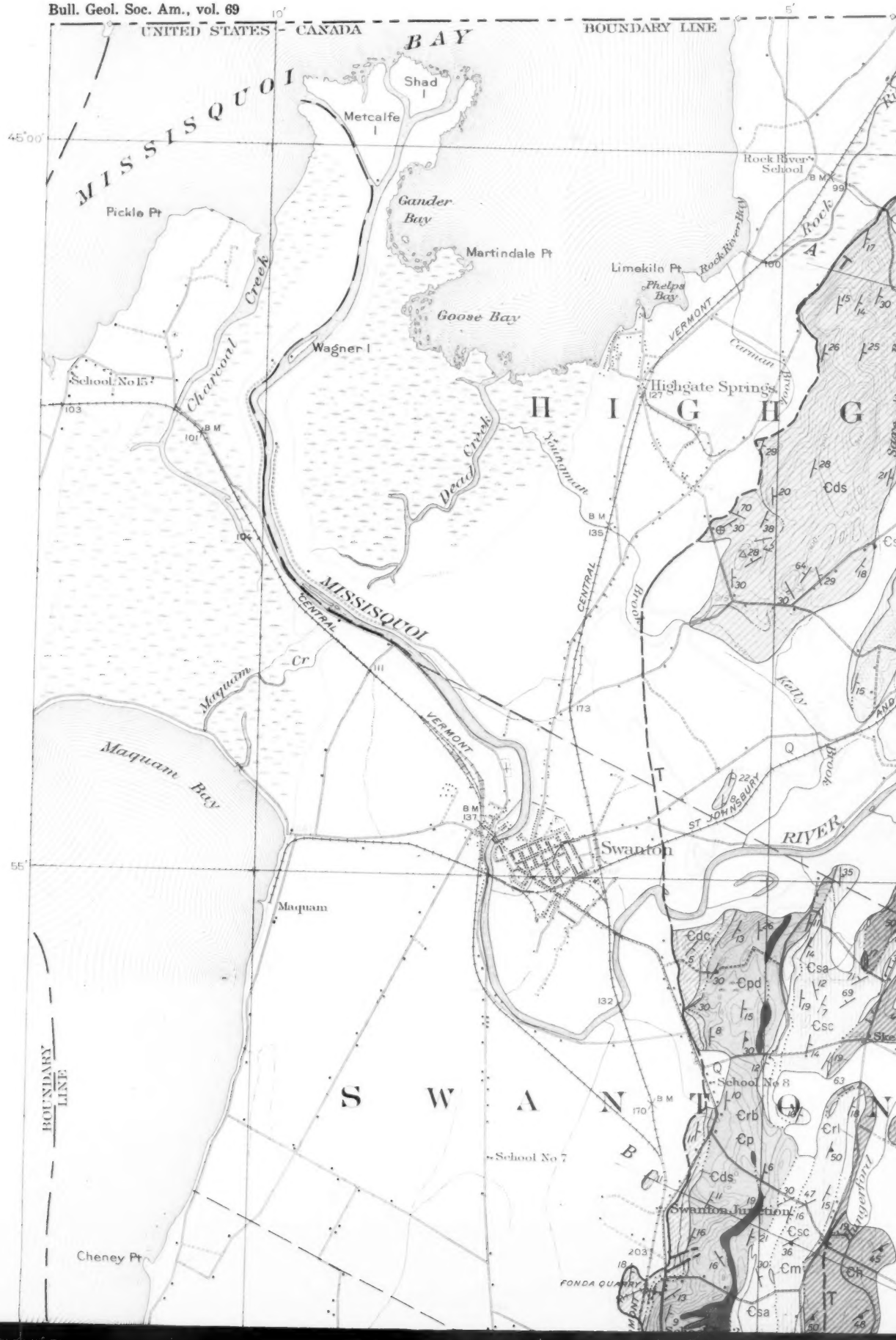
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Wagner I
Charcoal Creek
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Rock
Highgate Springs
VERMONT
Canman Brook
H I G H G
Dead Creek
Jamaman
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CENTRAL
MISSISQUOI
Maquam Cr
Maquam Bay
Kelly Brook
ST JOHNSBURY
RIVER
Swanton
Maquam
Cdc
Csa
Cpd
Csc
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School No 7
Swanton Junction
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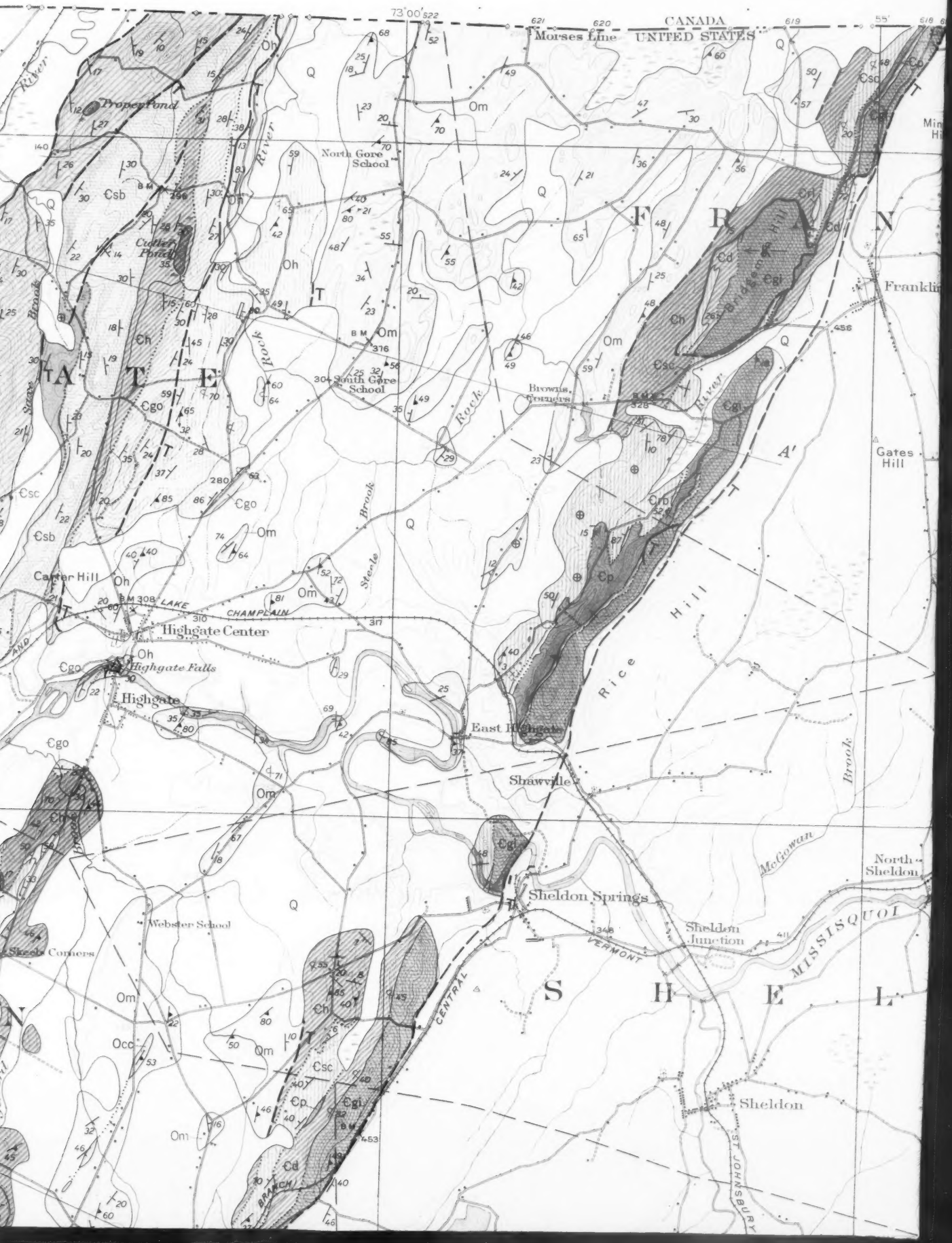
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Bridge Hill
Rice Hill
Carter Hill
Highgate Center
Highgate Falls
Highgate
East Highgate
Shawville
Sheldon Springs
Sheldon Junction
North Sheldon
Sheldon
Webster School
Skeels Corners
Rock River
Steele Brook
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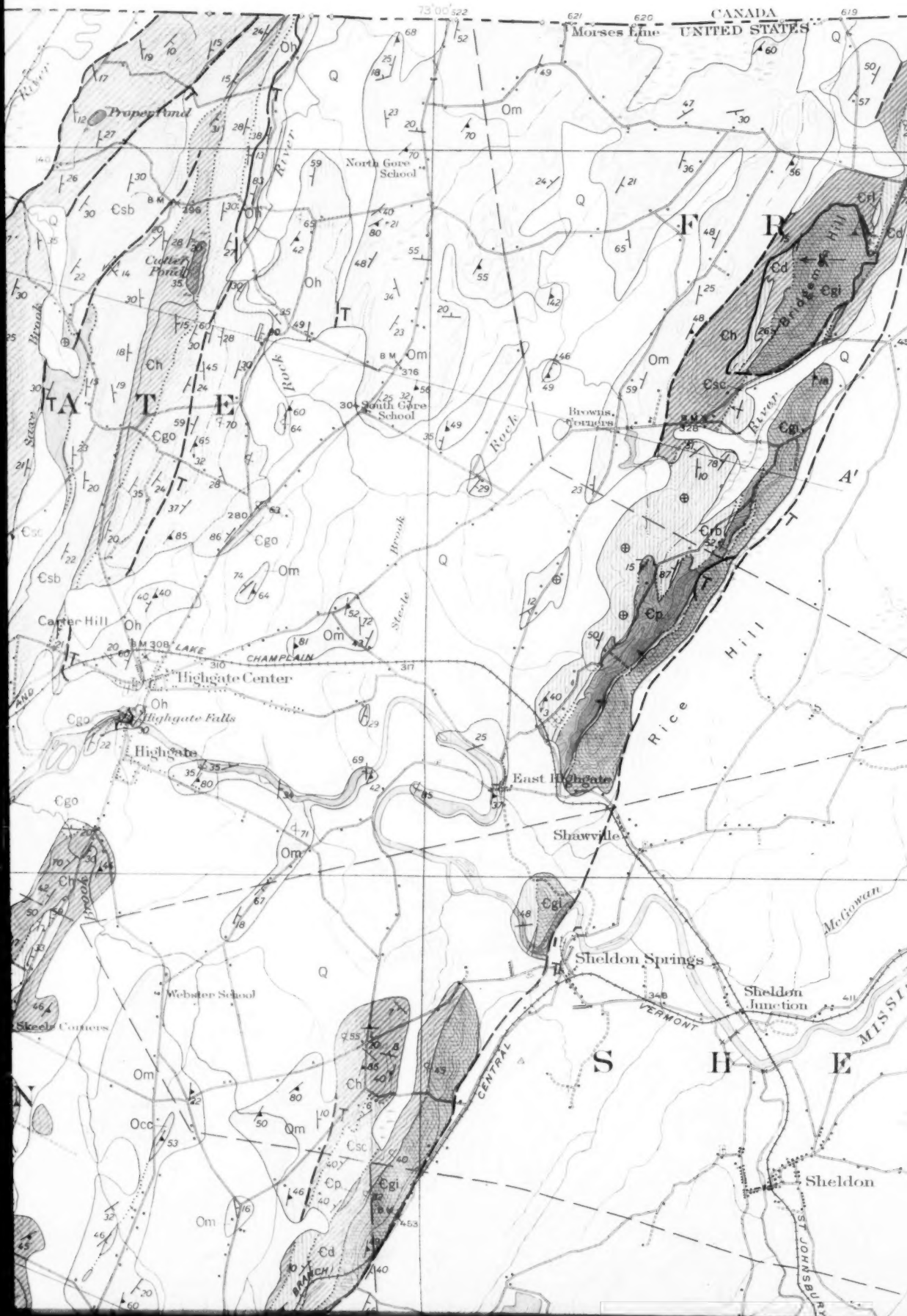
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Shawville
Sheldon Springs
Sheldon Junction
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Skeels Corners
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Central Vermont
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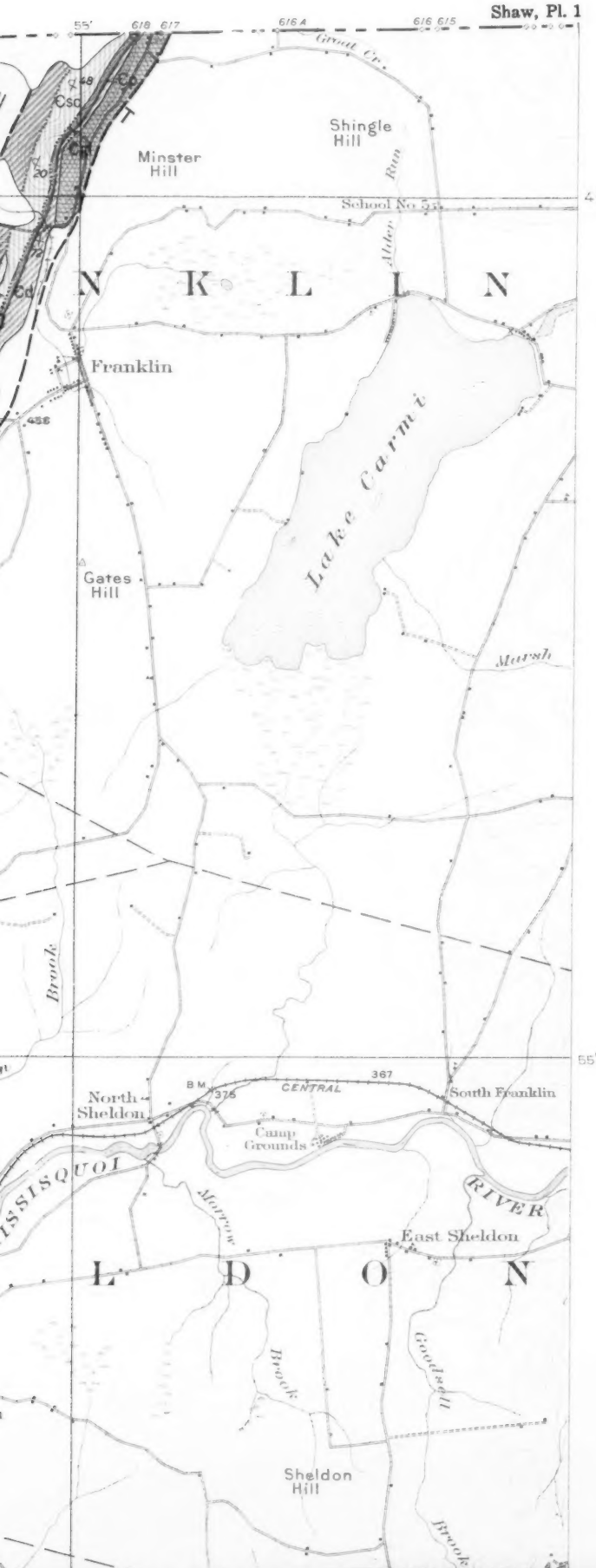

LEGEND

QUATERNARY

Q

Alluvium, beach deposits, moraine

MIDDLE (?) ORDOVICAN

Occ Om

Morses Line slate
(Occ = *Corliss conglomerate member*)

LOWER ORDOVICAN

Oh

Highgate formation

UPPER CAMBRIAN

Ɛgo

Gorge formation

MIDDLE CAMBRIAN

Ɛh

Hungerford slate

Ɛsb

Saxe Brook dolomite

Ɛrl

Rockledge conglomerate

Ɛsc

Skeels Corners slate

Ɛm

Mill River conglomerate

Ɛsa

St. Albans slate

Ɛrb

Rugg Brook dolomite

LOWER AND MIDDLE CAMBRIAN

Ɛp

Parker slate
(Ɛpd = *Middle dolomite member*)

Ɛd Ɛds Ɛdc

Dunham dolomite
(Ɛd = *Dunham undifferentiated*
Ɛds = *Sandy facies*
Ɛdc = *Carbonate facies*)

73°00'
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CANADA
UNITED STATES
Morses line
Proctor Pond
North Gore School
South Gore School
Browns Corners
Highgate Center
Highgate Falls
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Shawville
Sheldon Springs
Sheldon Junction
Sheldon
Webster School
Center Hill
LAKE CHAMPLAIN
Rock River
Rock Brook
Steele Brook
Rice Hill
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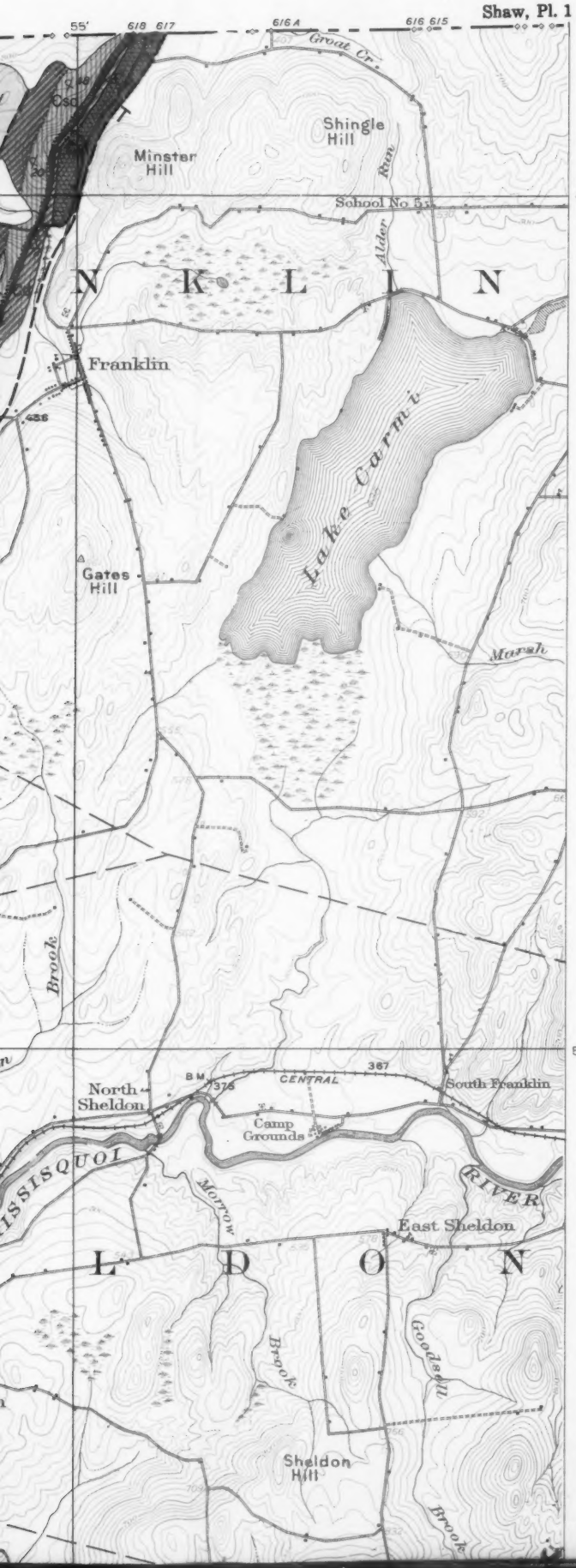


LEGEND

QUATERNARY

Q

Alluvium, beach deposits, moraine

MIDDLE (?) ORDOVICAN

Occ Om

Morses Line slate
(Occ = *Corliss conglomerate member*)

LOWER ORDOVICAN

Oh

Highgate formation

UPPER CAMBRIAN

Cgo

Gorge formation

MIDDLE CAMBRIAN

Ch

Hungerford slate

Csb

Saxe Brook dolomite

Crl

Rockledge conglomerate

Csc

Skeels Corners slate

Cm

Mill River conglomerate

Csa

St. Albans slate

Crb

Rugg Brook dolomite

LOWER AND MIDDLE CAMBRIAN

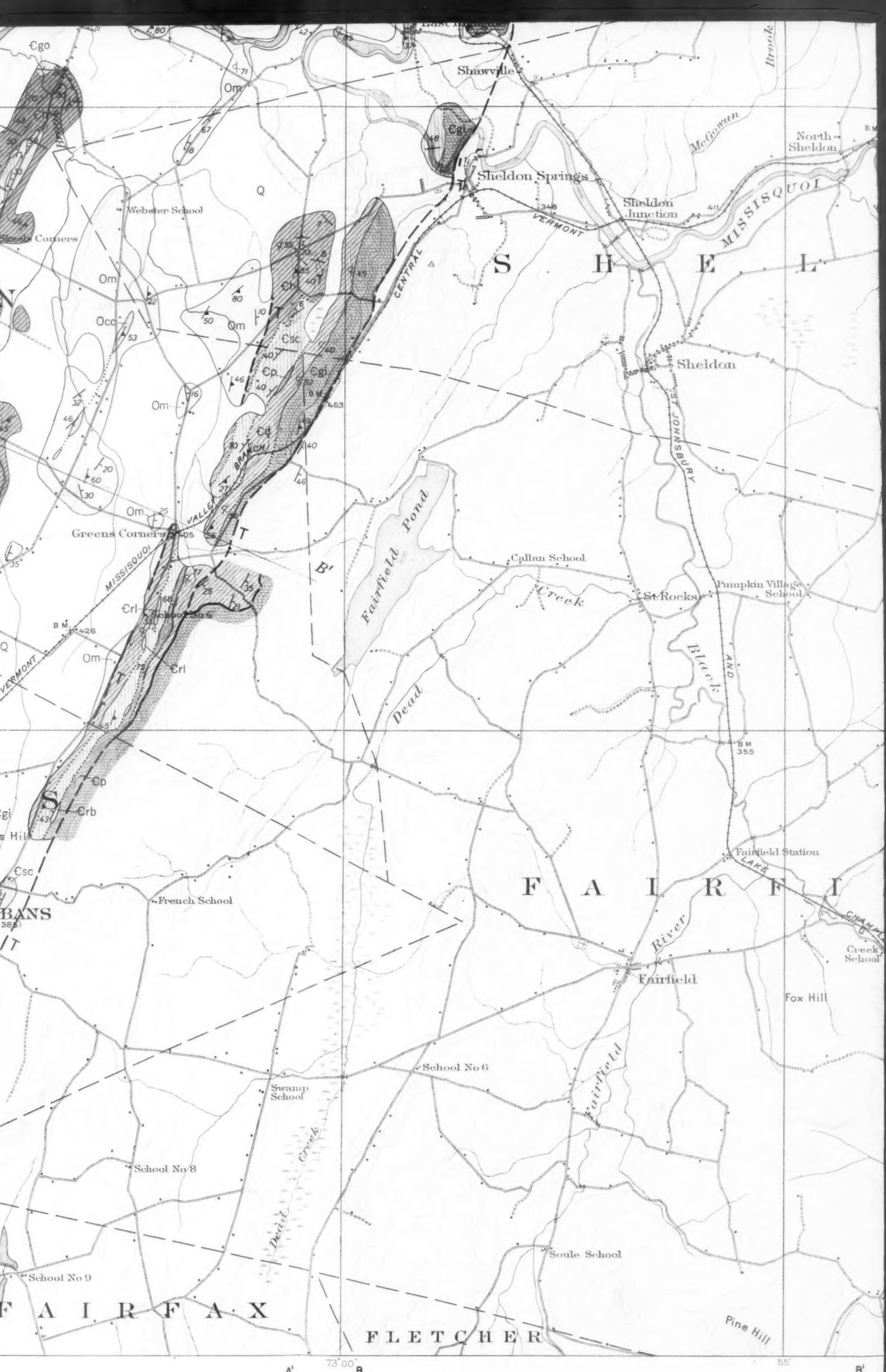
Cp

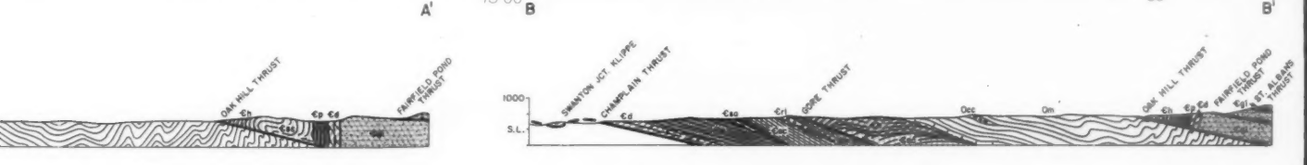
Parker slate
(Cpd = *Middle dolomite member*)

Cd Cds Cdc

Dunham dolomite
(Cd = *Dunham undifferentiated*
Cds = *Sandy facies*
Cdc = *Carbonate facies*)

GEOLOGIC MAP AND CROS





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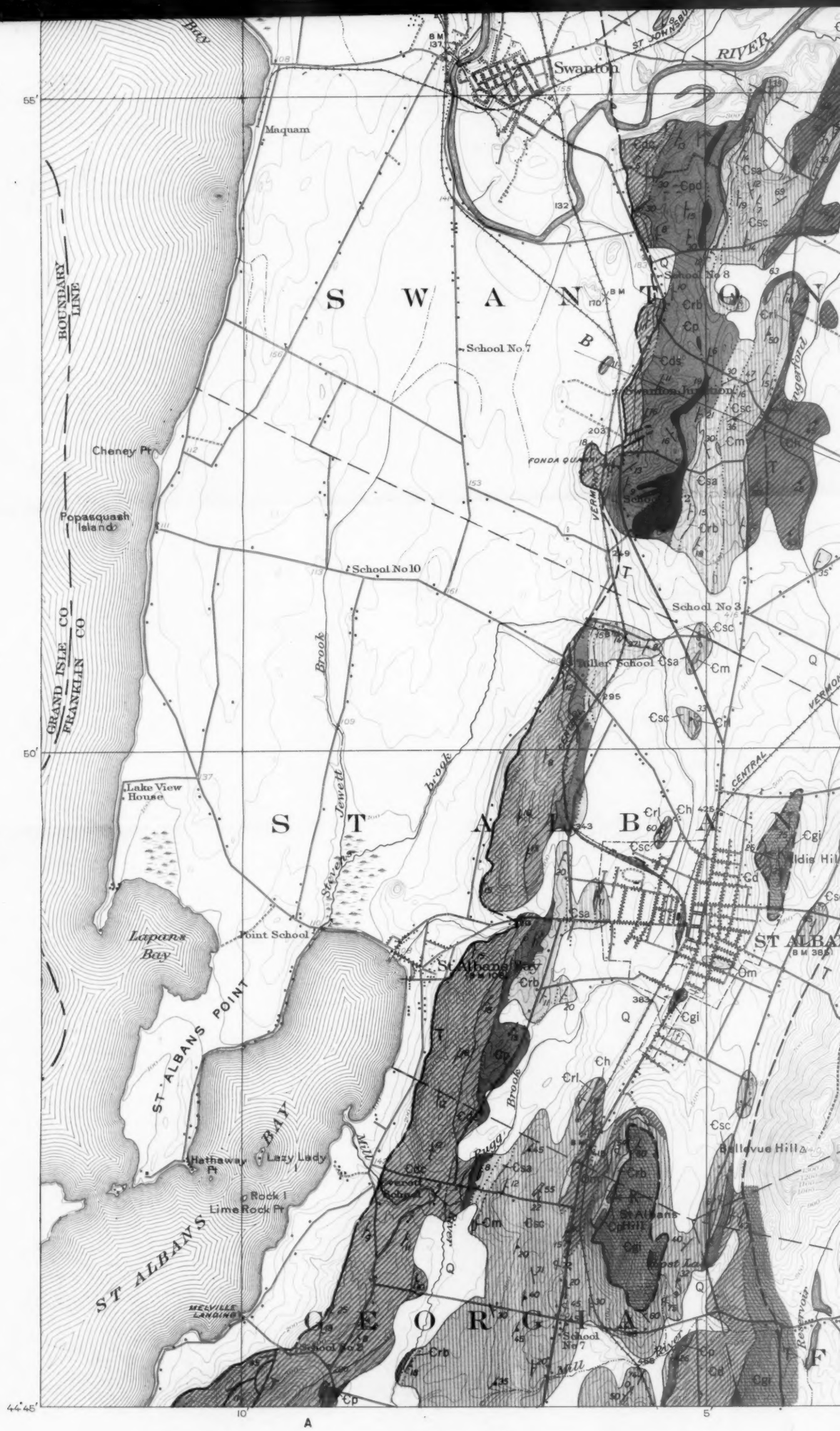
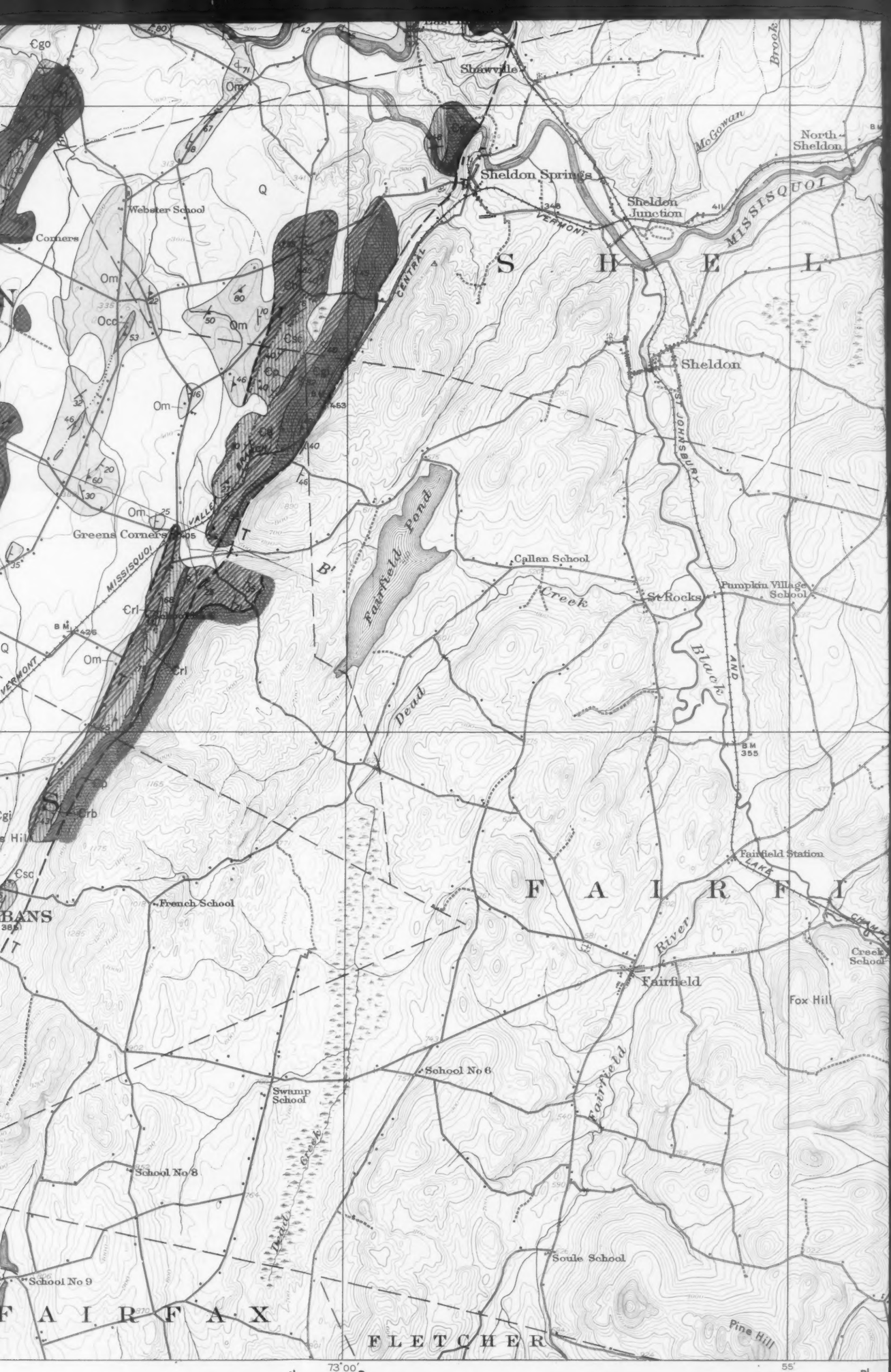


GEOLOGIC MAP AND CROS



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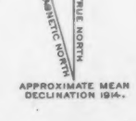
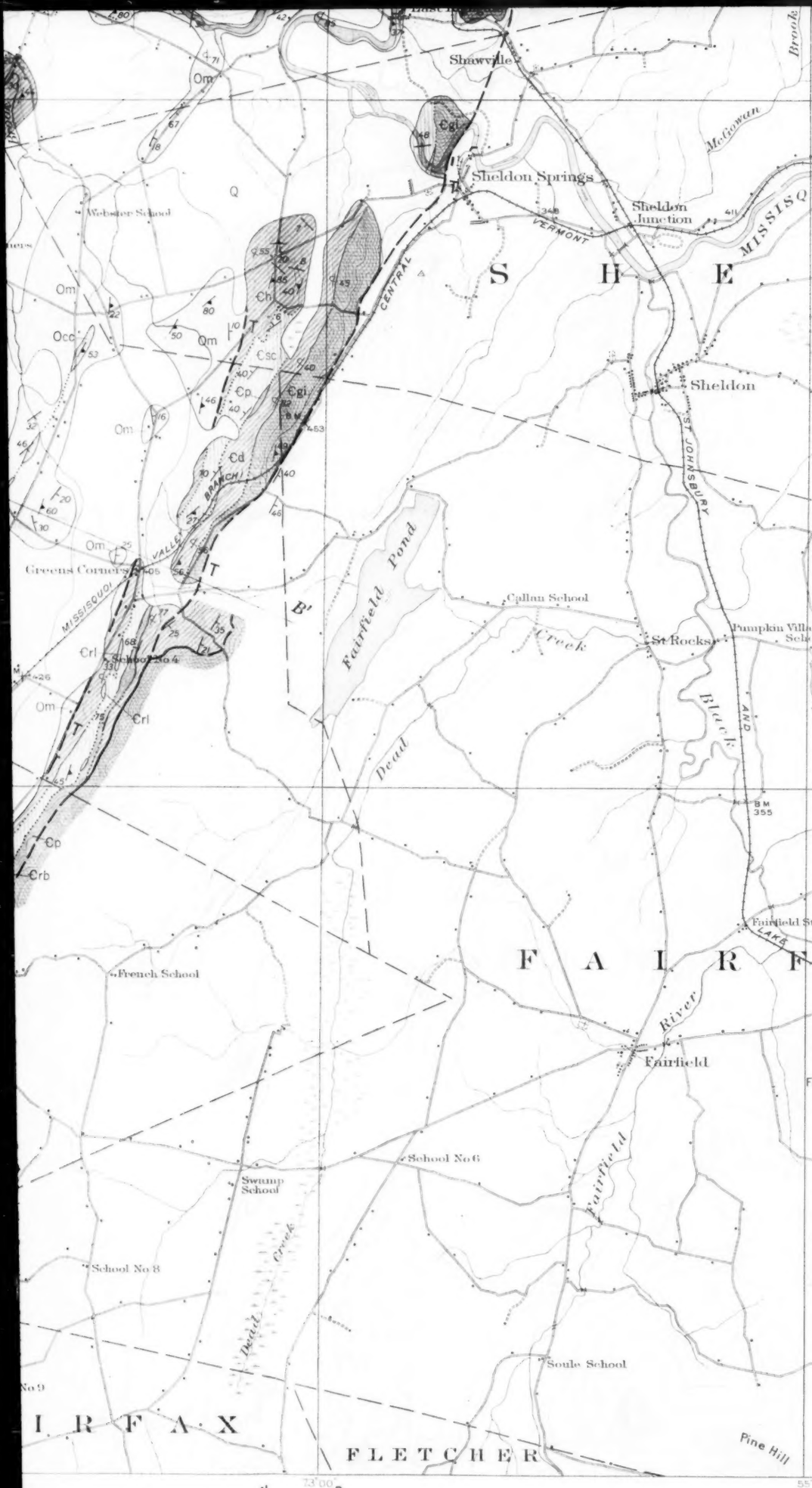




SECTIONS OF THE ST. ALBANS AREA, VERMONT

Scale $\frac{1}{62500}$

1 2 3 4 Miles

1 2 3 4 5 Kilometers

Contour interval 20 feet.

Datum is mean sea level.

Skeels Corners slate

$\mathcal{C}m$

Mill River conglomerate

$\mathcal{C}sa$

St. Albans slate

$\mathcal{C}rb$

Rugg Brook dolomite

LOWER AND MIDDLE CAMBRIAN

$\mathcal{C}p$

Parker slate
($\mathcal{C}pd$ = *Middle dolomite member*)

$\mathcal{C}d$ $\mathcal{C}ds$ $\mathcal{C}dc$

Dunham dolomite
($\mathcal{C}d$ = *Dunham undifferentiated*
$\mathcal{C}ds$ = *Sandy facies*
$\mathcal{C}dc$ = *Carbonate facies*)

$\mathcal{C}gi$

Gilman quartzite

SYMBOLS

Dip and strike of beds

Vertical beds with top toward west

Overturned beds

Horizontal bedding

Dip and strike of axial plane cleavage

Vertical axial plane cleavage

Dip and strike of axial plane cleavage and plunge of trace of bedding on cleavage

Dip and strike of axial plane of drag

Formation contact

Inferred formation contact

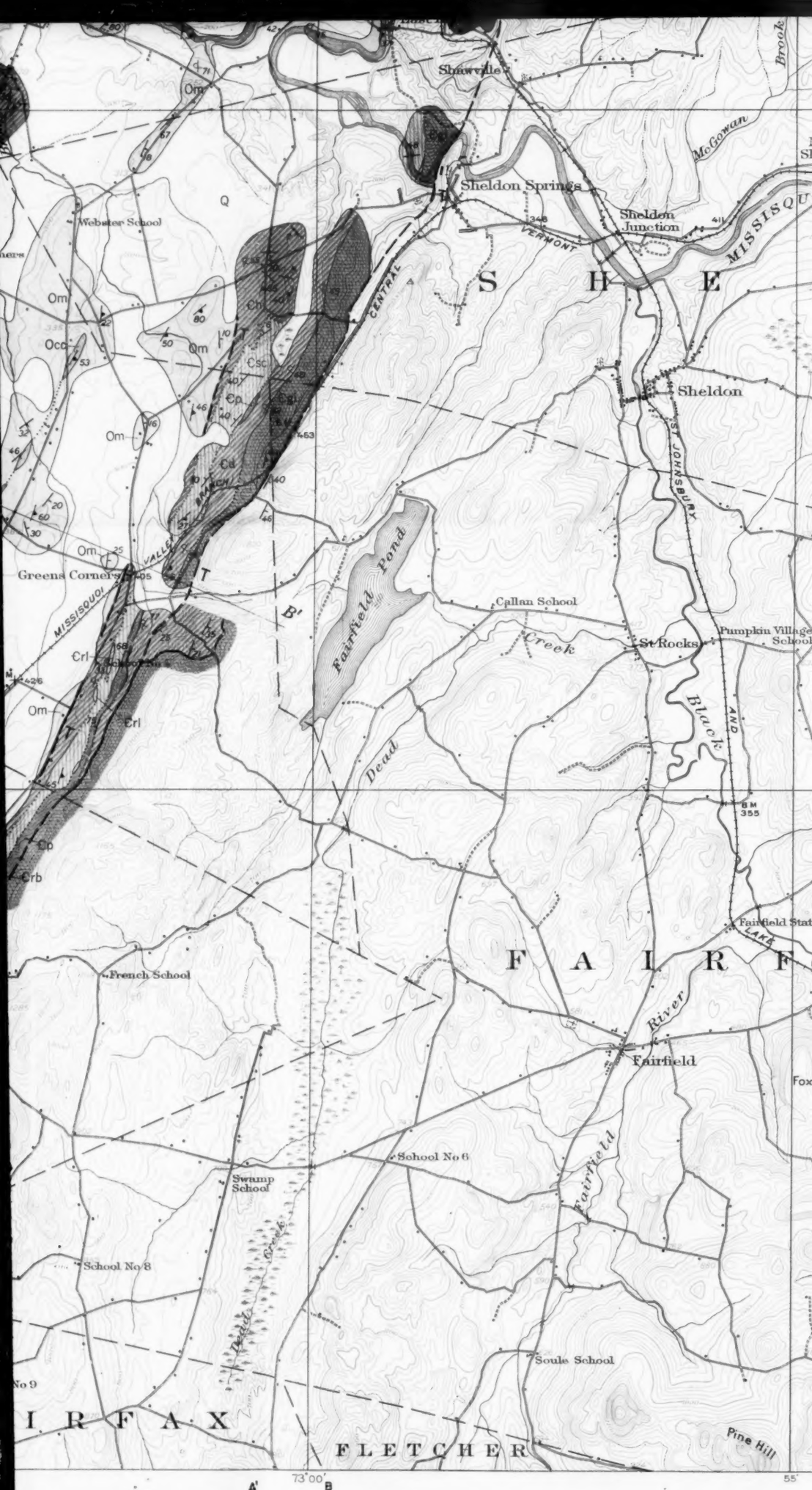
Fault

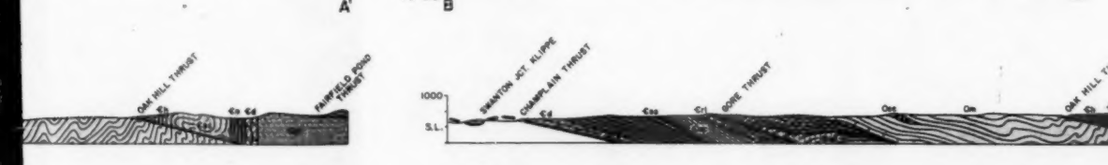
Inferred fault

K

Klippe, showing direction of movement

B' WILLIAMS & HEINTZ LITHOGRAPH CORPORATION, WASH. D. C.





SECTIONS OF THE ST. ALBANS AREA, VERMONT

Scale 1/62500

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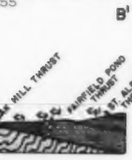
1 2 3 4 5 Kilometers

Contour interval 20 feet.

Datum is mean sea level.

WILLIAMS & HEINTZ LITHOGRAPH CORPORATION, WASH. D. C.



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Mill River conglomerate

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St. Albans slate

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Cgi

Gilman quartzite

SYMBOLS

Dip and strike of beds

Vertical beds with top toward west

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Horizontal bedding

Dip and strike of axial plane cleava

Vertical axial plane cleavage

Dip and strike of axial plane cleav
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STRATIGRAPHY AND STRUCTURE OF THE ST. ALBANS AREA, NORTHWESTERN VERMONT

BY ALAN B. SHAW

ABSTRACT

The Cambro-Ordovician sequence in northwestern Vermont includes two Lower Cambrian formations, Dunham Dolomite and Gilman Quartzite; Parker Slate is both Lower and Middle Cambrian, and Rugg Brook Dolomite is Middle Cambrian. Six formations of the Woods Corners Group (new name) are tentatively assigned to the upper Middle Cambrian because accepted evidence demonstrates that the *Cedaria* zone (lower Dresbachian) correlates with the standard Middle Cambrian of Europe rather than with the Upper Cambrian. The Upper Cambrian (post-Dresbachian), Lower Ordovician, and Middle (?) Ordovician are each represented by one formation: Gorge and Highgate formations and Morses Line Slate, respectively. The old names Georgia Slate, Mallett Formation, Winooski Dolomite, Russell Slate and Grandge Slate are abandoned; the Corliss Conglomerate is regarded as a member of the Morses Line Slate, and nearly all formations are substantially redefined. The faunas of each formation are listed and reviewed.

Structurally, the St. Albans area, which is basically a large northward-plunging syncline, is part of the Rosenberg thrust slice of Clark. East of the Champlain thrust six additional thrusts (four newly described) are recognized and cut the northern part of the area into five imbricated thrust slices. Minor folding and the flow cleavage are related to the major syncline rather than to the thrusting, although it is believed that the thrusting was a late stage in the same period of deformation that began with the synclinal folding. One tear fault seems to be genetically related to the minor folding.

Eight interformational unconformities and at least as many intraformational unconformities, particularly in the Ordovician, point out the structural instability of the St. Albans area in Cambro-Ordovician time. The predominance of slate contrasts to the carbonate-quartzite sequence to the south in the Hinesburg and Middlebury synclinoria. Sedimentary evidence suggests both northwestern and southwestern sources for the sands in the area.

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Introduction

The rocks of northwestern Vermont, in the vicinity of St. Albans, have been studied by geologists for the past century and a quarter, but neither a detailed stratigraphic column or geologic map has thus far been published.

The St. Albans area, as the term is used in this report, is a region of about 100 square miles in the northwestern part of Vermont (Fig. 1). Its northern limit is the International Boundary, and the southern limit is 44°45′ N. Lat., the southern edge of the St. Albans quadrangle map. On the east and west the area is bounded by thrust faults (Pl. 1). Thus defined, the St. Albans area is a strip about 18 miles long, from north to south, and 3–7 miles wide. Most of the area lies in the St. Albans quadrangle, but the northeastern corner lies in the Enosburg Falls quadrangle, which adjoins the St. Albans quadrangle on the east.

Local studies were made during July, August, and October, 1942, but systematic mapping of the area was carried on through the summers of 1946, 1947, and 1948. About 11 months were spent in the field. All field mapping was done on photostatic enlargements of the St. Albans and Enosburg Falls quadrangle maps, at a scale of 3 inches to the mile, except for small areas where pace-and-compass mapping was employed on a larger scale to clarify complex structural relations.

Because of deformation and abrupt lateral lithologic changes, each formation was walked out separately, in so far as glacial drift and other covering deposits allowed. The method of making closely spaced traverses across the strike and correlating between them was found to be unworkable, for sections only a few hundred paces apart are commonly so different that it is impossible to make positive correlations between them. Because of extensive glacial float, care was taken to collect fossils only from rocks in place as the beds were traced along the outcrop.

The thickness and ubiquity of the covering deposits (Pl. 1) made it impossible to measure detailed stratigraphic sections in the conventional manner. Therefore, all thicknesses given in this paper and those shown on Figure 3 have been computed trigonometrically from average dips and the width of outcrop, corrected for terrain, and should be regarded only as reasonable approximations. After the establishment of the stratigraphic succession, it would be profitable to study each formation in the area separately and to measure sections within each formation, but the present study was not made in such detail; its primary purpose was to determine the stratigraphic sequence and the principal structural features of the region.

Acknowledgments

The author is deeply indebted to Prof. Marland P. Billings, of Harvard University, who directed the field work and supervised the preparation of the doctoral dissertation, of

which this report is a revision; in addition, he read and criticized the revised manuscript before it was submitted for publication and permitted the use of a plane-table map made under his supervision as part of Figure 9 of this paper.

Dr. P. E. Cloud, Jr., Prof. R. R. Shrock, and the late Prof. P. E. Raymond gave many helpful suggestions in the field and laboratory, and Prof. B. F. Howell, of Princeton University, has aided the writer greatly by his critical reading of the completed thesis, by lending specimens from the paleontological collections at Princeton, and by answering many questions. Dr. D. L. Blackstone, Jr., of the University of Wyoming, read and criticized a preliminary draft of the manuscript submitted for publication. The writer is also indebted to two of his fellow students, Dr. Phillip T. Fowler, who assisted in the field in 1946, and Dr. S. W. Stone, who has mapped the Milton quadrangle, which lies south of the St. Albans area.

Dr. V. H. Booth, Dr. W. M. Cady, Dr. G. M. Kay, Dr. Franco Rasetti, and Mr. G. A. Sanderson have visited in the field and have given the author the benefit of their assistance and opinions. The people of Franklin County have been kind and helpful in every way. Mr. James C. Davis, Mr. and Mrs. Raymond McGinn, Mr. and Mrs. Joseph Daley, and Messrs. J. M. Boucher, John Pearo, and Norman Pelkey, all of St. Albans, and Mr. H. A. Roy, of Swanton, assisted in many ways that made the completion of the survey much easier. In addition, a portion of the field expenses during the summer of 1948 was defrayed by a grant-in-aid from the Sayles Fund of Harvard University.

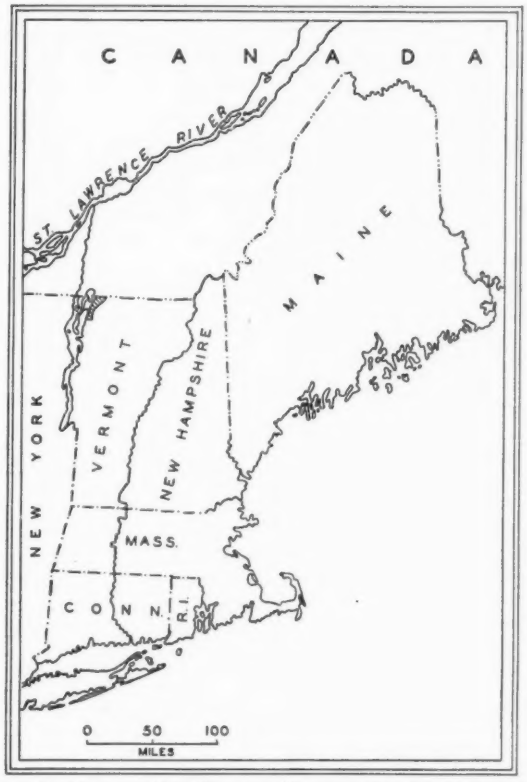

FIGURE 1.—MAP OF NEW ENGLAND SHOWING THE ST. ALBANS AREA (RULED)

Most recently, Dr. Cady and Dr. J. L. Wilson have acted as critics for The Geological Society of America and made numerous suggestions for improvement. The Shell Oil Company has assisted in preparation of the final manuscript and has given permission for publication.

Historical Summary

An extended account of the previous work done in and around St. Albans is omitted, for it would necessarily involve a discussion of the Taconic controversy, which is beyond the scope of this report. Therefore, only those works that have made some important contribution to the understanding of the local geology, or which are historically interesting are mentioned. Summaries of the development of geologic thought about the St. Albans-Georgia Center area have been given by Howell (1929) and Schuchert (1937, p. 1005–1011).

The first comprehensive study of the geology of Vermont, made by Edward Hitchcock and others (1862), is now of little more than historical interest. Sir William Logan, first Director of the Geological Survey of Canada, was the first to explain correctly the structure of northwestern Vermont. His investigations led him southward from Quebec to St. Albans, and he recognized the importance of the "great fault" now known as the Champlain overthrust (Logan, 1863, p. 281, 855, 859). Elkanah Billings (1861), Logan's able paleontologist, described several of the commonest Lower Cambrian fossils of the region and correctly interpreted the stratigraphic position of *Olenellus thompsoni*, which Hall (1859) had referred to post-Cambrian rocks.

During the two decades after Logan's work many authors discussed the rocks in Vermont directly or indirectly, but none of them contributed significantly to the geology of the St. Albans area (see the review by Walcott, 1891a, p. 91–114). The next important studies in northwestern Vermont were those of Walcott (1886; 1891b), who described and illustrated the fauna of the Lower Cambrian rocks, principally the Parker Slate.

Modern work in west-central and northwestern Vermont began with the studies of Keith (1923). He noted thrust faults east of the Champlain thrust, and he described and named many of the stratigraphic units now recognized. His paper also included the announcement of the discovery of Upper Cambrian fossils, described a short time later by Raymond (1924; 1925). Keith (1932) later revised his ideas and presented them in a second paper. After publication of Keith and Raymond's first papers, Howell (1926; 1929) formally named and described the Middle Cambrian St. Albans Slate, whose presence had been made known earlier by Perkins (1908, p. 209).

Schuchert (1933; 1937) revised the stratigraphy of northwestern Vermont in two papers that stood as the standard interpretation (Howell *et al.*, 1944) until Cady's work appeared in 1945. Schuchert's second paper was accompanied by paleontological contributions by Raymond and Howell.

Cady (1945) revised the Lower Cambrian succession and corrected many of the earlier misconceptions, but the area mapped did not extend into the St. Albans quadrangle. Most recently, the area to the east of that described in this report was mapped by Booth (1950). The region immediately north of the border was discussed briefly by McGerrigle (1931) and Clark (1934; 1936). The area to the west has been recently described by Hawley (1957).

While this article was in galley proof the description of the Highgate Springs sequence, which lies west of the Champlain thrust opposite the northern two-thirds of the St. Albans area, appeared in print (Kay, 1958).

Summary of Stratigraphy and Structure

The sedimentary rocks of the St. Albans area are, except for Quaternary deposits, entirely Cambrian and Ordovician (Table 1). Three camptonite dikes cut the Ordovician. At the base of the section are more than 4000 feet of Lower Cambrian clastic and calcareous rocks. The Middle Cambrian, most of which is slate, is nearly 3000 feet thick. The Upper Cambrian includes several hundred feet of Franconian and possibly Trempealeauan limestones and slates. The Lower Ordovician is about 3000 feet thick and is predominantly bedded limestone at the base, grading upward into calcareous shale. The Middle (?) Ordovician is represented by as much as 2500 feet of gray slate with intercalated

TABLE 1.—ROSENBERG SUCCESSION OF NORTHWESTERN VERMONT

Age	Thickness (feet)
Quaternary	
Lake sands, gravel, beach deposits and ground moraine; alluvium	0–200
Middle (?) Ordovician	
Morse Line Slate (including Corliss Conglomerate member)	0–2500+
Contact Unknown	
Lower Ordovician	
Highgate Formation	0–3000
Erosional Unconformity	
Upper Cambrian	
Gorge Formation	0–900+
Erosional Unconformity	
Middle Cambrian	
Woods Corners Group	
Hungerford Slate	25–400+
Saxe Brook Dolomite } equivalents	0–700
Rockledge Conglomerate } equivalents	0–100
Angular Unconformity	
Skeels Corners Slate	0–1200+
Mill River Conglomerate	0–35
Disconformity	
St. Albans Slate	0–200
Disconformity	
Rugg Brook Dolomite	0–500
Erosional Unconformity	
Lower and Middle Cambrian	
Parker Slate	0–1000+
Regional Unconformity	
Lower Cambrian	
Dunham Dolomite	25–2800+
Gilman Quartzite	500+

lenses of limestone conglomerate. Angular or erosional unconformities are present between the Middle and Upper Cambrian between or within several of the formations, and an erosional unconformity lies between the Cambrian and the Ordovician Systems.

Structurally, the St. Albans area is part of the large overthrust sheet called the Rosenberg slice (Clark, 1934), which has been folded into an asymmetric, northward-plunging syncline (Fig. 2). The western limb of the syncline dips gently to the east, but the eastern limb is vertical or overturned and has been pushed westward along the Oak Hill thrust so that it lies over the axis of the syncline. The eastern limb of the syncline is further modified by the Fairfield Pond and St. Albans thrusts, which, with the Oak Hill thrust, form the narrow Greens Corners slice. In the northern half of the area the Rosenberg slice is divided by the Saxe, Highgate Falls, and Gore thrusts into minor slices.

LOWER CAMBRIAN (GEORGIAN SERIES)

Gilman Quartzite

NAME: The name Gilman Quartzite was proposed by Clark (1934, p. 9) for a thick quartzite section in the Oak Hill succession in southern Quebec. It was introduced into the St. Albans area by Booth (1950, p. 1136, 1137, 1148, *etc.*), who described its outcrops in the Greens Corners slice, along the eastern side of the Rosenberg syncline (Pl. 1). This unit was also called the Brigham Hill Greywacke by Jacobs (1935, p. 85), but this name has not been generally adopted. Clark (1936, p. 144–146) and Booth have supplied additional details on facies development.

The Gilman is the stratigraphic equivalent of the Cheshire Quartzite of southern and central Vermont, but the Cheshire contains much clean white quartzite, and the argillaceous quartzites of the Gilman type are found only at at its base (Cady, 1945, p. 528). Because of this difference, the two names are retained.

DISTRIBUTION: The Gilman is present only along the eastern border of the mapped area, in the Greens Corners slice, and in the four klippen that lie on the Rosenberg syncline (Fig. 2; Pl. 1). These exposures have been partly mapped by Booth (1950). The resistant nature of the Gilman Quartzite causes it to form prominent outcrops wherever it reaches the surface; it accounts for the persistence of the four klippen mentioned above.

LITHOLOGY: Near St. Albans the Gilman is typically a pale-green, tightly cemented chloritic quartzite. Booth (1950, p. 1148) has

distinguished four lithic types in the formation, and it appears that the uppermost, which he describes as "a whitish relatively pure massive quartzite", comprises most of the exposures in the St. Albans area. Minor amounts of the second and third types, described, respectively, as "a fine-grained argillaceous siltstone" and "a mottled argillaceous quartzite" crop out on the eastern side of the St. Albans Hill.

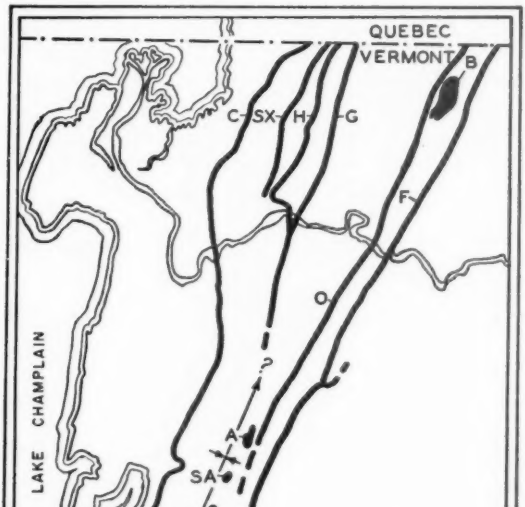

FIGURE 2.—GENERALIZED MAP OF THE MAJOR STRUCTURAL FEATURES OF THE ST. ALBANS AREA

C = Champlain thrust. *SX* = Saxe thrust. *H* = Highgate Falls thrust. *G* = Gore thrust. *O* = Oak Hill thrust. *F* = Fairfield Pond thrust. *S* = St. Albans thrust. The klippen are shown in solid black: *B* = Bridgeman klippe. *A* = Aldis klippe. *SA* = St. Albans klippe. *J* = Johnnycake klippe.

The point at which the main synclinal axis is overridden by the Oak Hill thrust is covered by glacial debris; for details *see* Plate 1.

THICKNESS: No measure of the total thickness of the Gilman is possible in the St. Albans area because it is everywhere faulted, but at least 500 feet are present. Cady (1945, p. 526) cites a thickness of 850 feet in west-central Vermont; Keith (1932, p. 396) gives the thickness at Wallingford, Vermont, as 800 feet, and Clark (1936, p. 146) estimated the formation to be at least 3000 feet thick at the type locality in southern Quebec, northeast of the St. Albans area. Booth (1950, p. 1149) says that in the Oak Hill succession (the sedimentary sequence east of the Fairfield Pond and St. Albans thrusts) in Vermont the Gilman is "certainly thicker than 850 feet given by Cady".

STRATIGRAPHIC RELATIONS: Only the upper part of the formation is exposed in the mapped area; the lower beds are cut off everywhere by thrust faults. At several places, especially northeast of Bridgeman Hill, the upper Gilman grades conformably into the overlying Dunham Dolomite through a 15- to 20-foot transitional zone of interbedded white to gray quartzite and brown-weathering dolomite. A 50-foot zone of transition has been noted to the south at the contact between the Cheshire Quartzite and the Dunham by Cady (1945, p. 526). This conformable relation of both the Cheshire and Gilman quartzites with the base of the Dunham affords strong evidence for the stratigraphic equivalence of the two quartzites.

FAUNA: The fauna of the Gilman is meager, but it is sufficient to date the unit as Lower Cambrian. Clark (1936, p. 146) recorded *Kutorgina* near Scottsmore, Quebec, and the

fauna in the St. Albans area includes *Salterella* (Shaw, 1954, p. 1034).

Dunham Dolomite

NAME: The rocks that are here assigned to the Dunham Dolomite have long been known in the St. Albans area as Winooski Marble (Keith, 1923, p. 109; 1932, p. 370; Schuchert, 1933, p. 361; 1937, p. 1023), but Cady (1945, p. 532) has shown that that name has been applied to two distinct units, and he has restricted it to the younger of the two, which is above the Monkton Quartzite, a formation in west-central Vermont. The Winooski as thus restricted is not present in the St. Albans area but is represented by its equivalent, the Rugg Brook Dolomite. For the older Winooski Cady has introduced Clark's term Dunham Dolomite from Quebec. This report marks the first use of the name in the St. Albans area save for local application in the Greens Corners slice by Booth (1950). Cady (1945, p. 528–529) has also reduced the old name Mallett from the rank of a formation to that of a member of the Dunham Dolomite. In this paper these rocks are not designated as a member but are simply referred to as the "sandy facies".

DISTRIBUTION: The Dunham Dolomite crops out along both the east and west sides of the mapped area and in two of the four klippen. It is a resistant unit, and wherever it is exposed it forms prominent ridges. On the west, along the Champlain thrust the Dunham formed the massive plate that carried the rocks of the Rosenberg slice westward along the Champlain thrust; here, the Dunham dips gently eastward and forms a low but persistent ridge that rises 100–250 feet above the level of the Ordovician foreland rocks to the west.

On the east side of the area the Dunham and the Gilman Quartzite together form prominent ridges along most of the length of the mapped area where the beds are vertical or overturned toward the west. Booth (1950) has mapped the Dunham at several places, but different interpretations of these areas are given on Figure 14 and Plate 1 of this report.

Along the western side of the Rosenberg syncline south of the Missisquoi River the sandy facies of the Dunham is confined to a thin zone at the top of the formation (the Mallett Member of Cady), but as McGerrigle (1931, p. 188) has noted, the carbonate facies is absent north of the river, and the entire 2800 feet of the Dunham belongs to the sandy facies (Fig. 3). The present study confirms Booth's observations (Booth, 1950, p. 1150–1151) that only Dunham carbonates are present on the eastern side of the area. Plate 1 and Figure 3 of this report show diagrammatically the facies distribution within the Dunham on the western side of the St. Albans area.

LITHOLOGY OF CARBONATE FACIES: On the west, the carbonate facies is exposed only south of the Missisquoi River, except for a single exposure in an old quarry 2.4 miles S. 70°W. of Highgate Center. This facies is developed southward beyond the limits of the mapped area, well into west-central Vermont (Cady, 1945, maps). The principal rock is a coarse-grained dolomite. The lower part of the facies consists of gray dolomite and minor amounts of red dolomite. Thin, resistant beds of dense, argillaceous gray dolomite stand out on weathered surfaces as prominent ridges. The weathered surfaces show many shades of brown, red brown and tan. Above these lower gray dolomites but not sharply demarcated from them lies the mottled red Wakefield, Swanton or calico marble. On fresh surfaces these dolomites range in color from light orange through red to violet; the weathered surfaces are smooth, in contrast to the differential weathering found in the lower beds, and they show many shades of brown. Breccia composed of angular fragments of red dolomite in a matrix of white dolomite is uncommon, and throughout the upper part of the facies there are thin wavy partings of bright-red to red-violet shale, which weather out as prominent, thin ridges (Schuchert, 1937, Pl. 4, fig. 1). Above the mottled dolomites lies another zone of gray dolomite, locally containing quartz sand. Sandstone, pure limestone, and shale are present in the carbonate facies, but they are rare.

On the eastern side of the area Dunham Dolomite seems to be mainly in a thick-bedded, differentially weathering phase that in part is believed to correlate with the lowest beds on the west, but which in large part seems to be older than the strata exposed above the

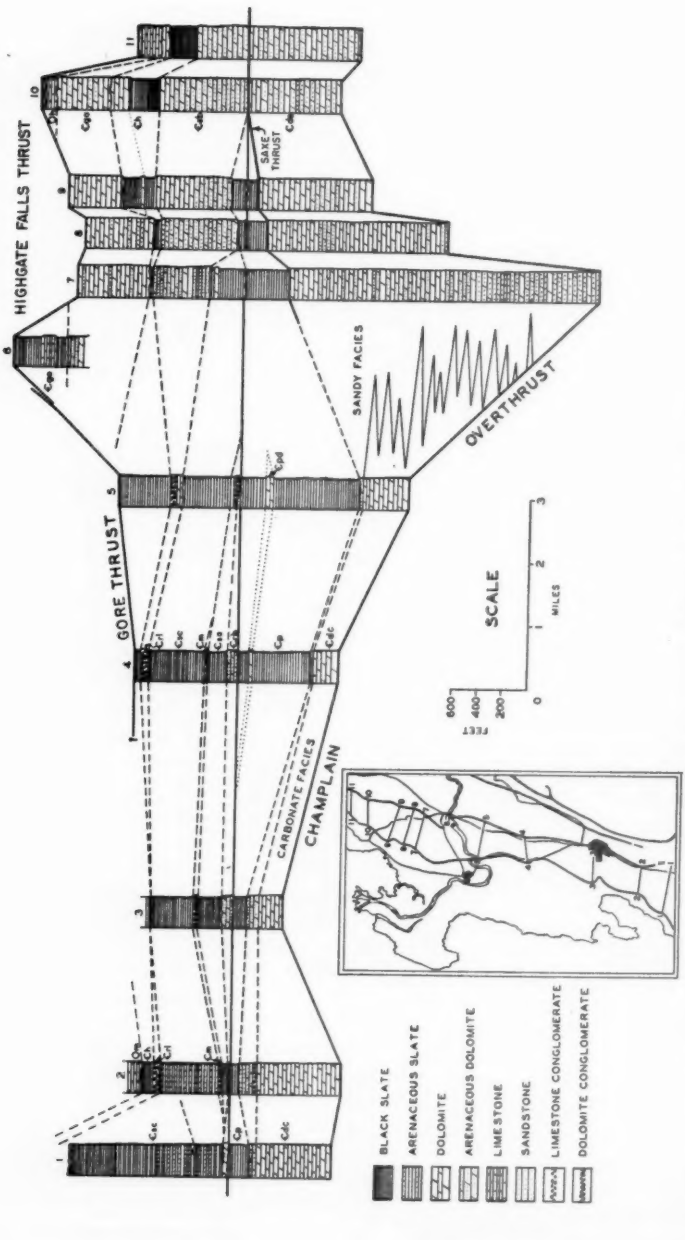

FIGURE 3.—COLUMNAR SECTIONS, ST. ALBANS AREA

Along the west side of the Rosenberg slice. Thickness of all units derived trigonometrically. €dc = Dunham Dolomite (carbonate facies). €ds = Dunham Dolomite (sandy facies). €p = Parker Slate. €pd = Middle Parker dolomite. €rb = Rugg Brook Dolomite. €sa = St. Albans Slate. €m = Mill River Conglomerate. €sc = Skeels Corners Slate. €rl = Rockledge Conglomerate. €sb = Saxe Brook Dolomite. €h = Hungerford Slate. €go = Gorge Formation. Oh = Highgate Formation. Om = Morses Line Slate.

Champlain thrust. These eastern dolomites have beds 4 inches–3 feet thick separated by more argillaceous interbeds up to 6 inches thick.

LITHOLOGY OF SANDY FACIES: South of the Missisquoi River the sandy facies is composed almost exclusively of gray dolomitic sandstone. It lies above the carbonate facies and averages 35 feet in thickness, but reaches a maximum of 150 feet. North of the river the sandy facies thickens and replaces the carbonates (Fig. 3), and in place of the uniform gray dolomitic sandstone there is a thick sequence of complexly interlaminated beds, tongues, and lenses of gray dolomite, gray sandy dolomite, white quartzite, red arkosic sandstone, black shale and other minor rock types.

There appear to be two general zones. The lower zone contains much sandstone and quartzite and some red arkosic sandstone. This zone is best developed northward from the topographic nose 1.08 miles S. 65° E. of Highgate Springs to the site of Saxe's Mill (where Saxe Brook flows westward across the Dunham). These are the beds that Schuchert (1937, p. 1028) and Keith (1923, p. 107) confused with the Monkton Quartzite; they change southward into gray sandstone and dolomite. Locally gray and black dolomitic shales lie near the bottom of the sandy facies and are exposed on the 340-foot hill 1.57 miles S. 33° E. of Highgate Springs.

A second zone lies essentially above the first but grades southward into the upper part of the first. Near the International Boundary the rocks in the upper zone are mainly gray calcareous sandstones, but southward there are many quartzite beds, some up to 15 feet thick. Still farther south the quartzites grade into less arenaceous gray dolomite. At the International Boundary the upper zone contains much black and dark-gray chert, which seems to be secondary for it forms small masses that cut across the bedding. This chert has been found in all Cambrian sandstones and carbonates in the northwestern corner of the mapped area, and apparently identical chert is also present as pebbles in the Upper Cambrian Saxe Brook Dolomite, near the International Boundary.

CONTACT BETWEEN FACIES: Howell (1934) noted the presence of a disconformity between the sandy and carbonate facies southwest of St. Albans, and he assumed that "there must have been small crustal movements in northwestern Vermont between the time of deposition of the uppermost Winooski (*i.e.*, the carbonate facies) and . . . the first Mallett (*i.e.*, sandy) sediments." A similar contact may be seen at the base of the outcrop 1.4 miles N. 70° W. of School No. 7, 3¾ miles south of St. Albans, where the sandy facies contains quartz pebbles 2–6 inches long. Also, the upper 8 feet of the sandy facies are conglomeratic behind Tuller School, 2½ miles northwest of St. Albans. However, breccias and conglomerates are present at many stratigraphic levels throughout the St. Albans area, and it would seem to be placing undue emphasis on these particular conglomerates to regard them as indicative of an important systemic break, for as Schuchert (1937, p. 1023) has pointed out, "The seas of those times were shallow, and intraformational conglomerates are not rare, hence, local breaks and such conglomerates are to be expected."

THICKNESS: The exposed thickness of the Dunham changes abruptly and irregularly along the west side of the St. Albans area because beds are cut out by the Champlain thrust. The thickest section (2800 feet) lies east of Highgate Springs (Fig. 3, col. 7).

Along the eastern margin of the area, and in the Greens Corners slice, the Dunham dwindles from a maximum thickness of more than 1500 feet at the southern boundary of the St. Albans quadrangle to about 25 feet at the International Boundary. This thinning is a reflection of the fact that the Dunham probably thins from west to east whereas the line of outcrop crosses the isopach lines diagonally from southwest to northeast (Fig. 4).

STRATIGRAPHIC RELATIONS: The contact between the Dunham and the overlying Parker Slate is an erosional unconformity. The contact between the Dunham Dolomite and the Gilman Quartzite is exposed only on the eastern limb of the Rosenberg syncline, in the Greens Corners slice, and there it is conformable, as has already been described.

FAUNA: The fauna of the Dunham Dolomite has been discussed in detail elsewhere (Shaw, 1954, p. 1036–1037; 1955b) and is listed below. The Dunham seems to be middle to late Early Cambrian.

The fossils whose presence in the Dunham seems to be reasonably well established include: Trilobita—*Antagmus typicalis*, *A.? simplex*, *Billingsaspis adamsii*, *Bonniella desiderata*, *Olenellus* sp. indet., *Perimetopus arenosus*, *Ptychoparella teucer*; Brachiopoda—*Nisusia festinata*; Gastropoda—"*Scenella*" *varians*; Hyolithida—*Hyolithes americanus*, *Hyolithes* sp. indet.; "Vermes"—*Planolites congregatus*.

Fossils that have been reported from the Dunham but whose presence has not been confirmed by recent study include: Trilobita—*Atops trilineata;* Brachiopoda—*Botsfordia caelata*, *Obolella crassa*, *Paterina labradorica swantonensis*, *Wimanella? orientalis*; Hyolithida—*Hyolithes communis*, *Hyolithellus micans*.

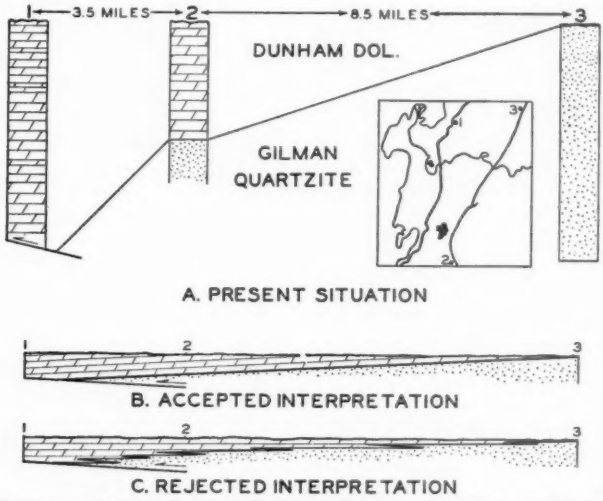

FIGURE 4.—INTERPRETATIONS OF CONTACT BETWEEN DUNHAM DOLOMITE AND GILMAN QUARTZITE
Distances between sections have been increased to compensate for shortening by later thrusts. *See* explanation in text.

LOWER AND MIDDLE CAMBRIAN

Parker Slate

NAME: This formation was named by Keith (1932, p. 371) "from its excellent exposures around the sides of the Parker Cobble and on the old Parker Farm", 2¼ miles N. 60° W. of Georgia Center, in the Milton quadrangle. The name replaced Colchester Formation (Keith, 1923, p. 110), an earlier but less satisfactory term.

This unit might properly be called the Georgia Slates, for Walcott's original Georgia Shales (Walcott, 1886, p. 15–17) began at Parker's ledge, the type locality of the Parker Slate, but the old name has not been retained because of its confused history. Walcott (1886) included the Upper Cambrian Skeels Corners Slate and Rockledge Conglomerate as Units 7–10 of the Georgia shales, and, later, Keith (1932, p. 378–379) applied the name Georgia Slate to a heterogeneous group of Upper Cambrian and Lower Ordovician beds now assigned to the Skeels Corners Slate, Hungerford Slate, and Highgate Formation. Cady (1945, map) has mapped but has not defined or described a group of Upper Cambrian slates and limestone conglomerates which is said to include the Georgia (?) Slate and two other formations of Dresbachian age. This is still another use of the term for Upper Cambrian rocks. In view of these manifold applications, the names Georgia Slate and Georgia Shale are not used in this report.

DISTRIBUTION: The Parker Slate included at its type locality the famous Noah Parker quarry, on Parker Cobble, west-northwest of Georgia Center in the Milton quadrangle, from which the first specimens of *Olenellus thompsoni*

were obtained and from which Walcott (1886; 1891b) made his extensive collections of the *Olenellus* fauna.

The Parker is most extensive along the western side of the St. Albans area (Pl. 1). South of the Missisquoi River it forms a lowland belt nearly a mile wide and extends southward to the latitude of Fonda Quarry, south of which the formation is concealed by glacial cover for a mile. South of Stevens Brook the Parker is again exposed, but in a band only about 1000 feet wide. The formation continues in this reduced outcrop for a few miles south of the southern edge of the mapped area and there disappears; it is replaced still farther south by the Monkton Quartzite.

Slate containing fossiliferous limestone conglomerate and limestone bioherms and lying above the Dunham Dolomite and below the Saxe Brook Dolomite north of the Missisquoi River is assigned to the Parker (Shaw, 1954, p. 1043–1044). Schuchert (1937, p. 1031) reported fossiliferous Lower Cambrian sandy limestone and slates "two miles North 15° West of Highgate Center" at some point "on the eastern side of Saxe Valley", but this outcrop could not be relocated.

On the eastern side of the area slate and arenaceous dolomite above the Dunham are referred to the Parker on the basis of their stratigraphic position. The three principal areas of outcrop are southeast of St. Albans, east of Greens Corners, and northward from the East Highgate railroad station to the International Boundary (Fig. 16).

LITHOLOGY: The Parker on the west is dominantly gray to black, abundantly micaceous slate or shale. The mica flakes may be original detritus. Arenaceous and non-arenaceous dolomite is present throughout the formation but is a minor constitutent except for one persistent zone near the middle of the formation (Pl. 1; Fig. 3), which has been designated the middle Parker dolomite (Shaw, 1954, p. 1038). Immediately below the middle dolomite, through a stratigraphic interval of 50–100 feet there is a zone of discontinuous dolomite bioherms. Locally, on the western side of the area and south of the Missisquoi River the arenaceous dolomites contain sandstone beds as, for example, north of the road east from Fonda Quarry and southeast of St. Albans where quartzite lenses are present near the base of the formation. Conglomerate is uncommon. Limestone bioherms are rare and all are confined to the formation north of Fonda Quarry and in Saxe Valley.

A ledge of limestone conglomerate in the Saxe Valley 2 miles N. 20° W. of Highgate Center contains an early Middle Cambrian (*Albertella* zone) fauna identical with that in the upper Parker Slate south of the Missisquoi River; for that reason it is assigned to the Parker Slate in spite of its different lithology (Shaw, 1954, p. 1043–1044; 1957).

On the eastern side of the St. Albans area slate outcrops are rare in the Parker. North of East Highgate the formation contains beds of quartzite and dolomitic sandstone which can be accurately traced. These beds were mapped by Booth (1950) as Rugg Brook Dolomite, which they do resemble, but this correlation does not seem to be correct because the sequence Dunham-Parker-Rugg Brook-Skeels Corners can be traced north from the knoll east of St. Albans Hill where the succession is clearly exposed and where fossils are present to identify the Skeels Corners.

An unusual lithic type is a graywacke consisting of fragments of slate and sandy dolomite, up to an inch and a half in greatest dimension, enclosed in a matrix of gray, slightly dolomitic shale, which crops out on a small hill 0.63 mile due east of the railroad grade crossing, a mile north of East Highgate.

THICKNESS: In the southwestern part of the St. Albans quadrangle the Parker Slate is probably about 250 feet thick, but little of it is exposed. North of Fonda Quarry the formation thickens to a maximum of 1000–1200 feet (Fig. 3). It is impossible to make precise measurements of thickness because of the folding, minor thrust faulting, and heavy glacial cover.

On the eastern side of the area it is difficult to determine the upper limits of the formation so that estimates of thickness are not reliable. From Greens Corners south the Parker is locally missing and does not appear to exceed 50 feet anywhere, but northeastward, there seem to be 200–250 feet of beds. It is not certain that the Parker crops out at the International Boundary, but it is present less than a mile to the south and appears to be represented to the northeast, in Canada, by the Oak Hill Slate

(Clark, 1936, p. 148). Booth (1950, p. 1136, 1151) has recognized the Parker as a mappable unit in the Oak Hill succession in Vermont.

STRATIGRAPHIC RELATIONS: Before the deposition of the Parker Slate the Dunham Dolomite was exposed to erosion. Channeling has been described by Schuchert (1937, p. 1025, 1034) at a locality less than a mile north of Swanton Junction, and the eroded contact is also exposed in a pasture immediately southeast of the Central Vermont Railroad tracks 0.62 mile S. 38° E. of the Tuller School, where arenaceous dolomite of the sandy facies of the Dunham is channeled to a depth of about 8 feet; here the Parker contains boulders and pebbles of the underlying dolomite embedded in a fossiliferous gray slate matrix. Schuchert (1937, p. 1025) believed that "the erosion interval was not a long one" between the deposition of the Dunham and the Parker, but mapping has brought out the fact that at least 2000 feet of Dunham are missing on the eastern side of the St. Albans area. Some control on the changes in thickness of the Dunham may be gained from sections on both sides of the syncline, and three selected sections have been projected into a single diagram in Figure 4*A*. The two possible interpretations of these sections are given in Figures 4*B* and 4*C*.

One interpretation is that the Gilman and Dunham are lateral equivalents (Fig. 4*C*). If this is so, the eastward thinning of the Dunham merely represents the replacement of the Dunham interval by the Gilman Quartzite, and the post-Dunham erosion surface would be indicative of but a short period of exposure.

Figure 4*B* represents what seems to be the more likely interpretation that the Gilman and the Dunham were deposited successively. Then, before the deposition of the Parker Slate, tilting exposed the Dunham to erosion, which was more vigorous to the east and removed nearly all of the Dunham. The depth of erosion could have been due either to differential uplift or to longer exposure; there is little evidence on this point at present. If the sheet of Dunham was originally about 12 miles wide it would require only 3° of tilting to raise the eastern side of the sheet 3000 feet above the western. Such a slight angular discordance between the Dunham and the Parker would hardly be discernible in the outcrop.

The Parker also thins toward the east much like the Dunham. This may be due either to repeated uplift in Middle Cambrian time or to the persistence of a shelf to the east after the erosion of the Dunham. The obvious evidences of severe Middle Cambrian erosion brought out by this study (Fig. 3 of this paper; Shaw, 1954, p. 1044, Fig. 4) favor the first interpretation, but they do not constitute proof.

FAUNA: The fossils of the Lower Cambrian in northwestern Vermont have been studied and discussed at great length, at times acrimoniously, since the discovery of *Olenellus thompsoni* on the Parker farm nearly a century ago. Billings (1961) early described most of the fossils from what is now known as the Dunham Dolomite, and the classic studies of Walcott (1886; 1891b) presented quite fully the Parker Slate fauna. In more specialized studies, Walcott (1910) and Resser and Howell (1938) have revised the olenellids and have figured many Vermont specimens. Resser (1937) refigured Billings' types and recognized several new species, some of which are of dubious value. The most recent works on faunas of this age are the report on the Monkton Quartzite fauna (Kindle and Tasch, 1948; Tasch, 1949) and the summaries of the Lower Cambrian faunas by Shaw (1954; 1955b). The most important recent addition to the knowledge of the Parker faunas was the identification of Middle Cambrian fossils in the upper part of the Parker.

The lower Parker fauna (Lower Cambrian) is essentially that described by Walcott (1886; 1891b) as the typical *Olenellus* zone fauna, dominated by such genera as *Olenellus*, *Bonnia*, seven-spined *Kootenia*, *Pagetides*, and the brachiopods *Nisusia* and *Kutorgina*. The upper Parker fauna (Middle Cambrian) is probably present as low as 580 feet above the base of the formation, beneath the middle Parker dolomite, but it is present in its fullest development approximately 190 feet above the middle Parker dolomite. Here the fauna is dominated by six-spined *Kootenia*, *Nisusia cloudi*, and such other characteristic *Albertella* zone fossils as *Mexicella*, *Oriella*, and *Zacanthoides*. This fauna has also been described (Shaw, 1957). The two faunas are listed in Table 2 (with changes from

Shaw, 1955b), but data on stratigraphic ranges and relative abundance of the species is given elsewhere (Shaw, 1954). The upper Parker slates are much more restricted areally than the lower, and, hence, the Middle Cambrian faunas are rarer.

Paleogeography

Stratigraphic differences between northwestern and west-central Vermont indicate that the St. Albans area was probably a basin during most of the Cambrian. It was partially restricted to the south by a positive area (Fig. 6). Streams carried clastics into the basin from the Adirondack-Laurentian landmass. The high toward the south caused concentration of clastics in northwestern Vermont as shown in Figure 6. Throughout Cambrian and Early Ordovician times the St. Albans area also seems to have been a region of considerable crustal unrest, as is evidenced by the many unconformities present in the column (Fig. 5). The continuity of these breaks into west-central Vermont is not certain.

The partial isolation of the area seems to have been due to a positive eastward-trending arch through Vermont between Milton and Burlington. North of this arch, which is here named the Milton High, lay the clastic-filled basin which the writer proposes to name Franklin Basin, since most of its known extent is in Franklin County, Vermont.

It seems that the first movements in the Franklin Basin took place between the end of Dunham time and the beginning of Parker deposition, as has been discussed above. The Milton High itself seems to have come into existence during the period of movement following the deposition of the Parker because the Dunham continues more or less unchanged across the high, whereas the Parker is eroded away toward the south and its southern equivalent, the Monkton Quartzite, is eroded away toward the north. The evidence that this complementary thinning is due to erosion and is not an original feature has been presented elsewhere (Shaw, 1954, p. 1044).

In reviewing this paper as a critic, J. L. Wilson asked whether the clastics of the St. Albans area might have been brought in from the east by thrusting more severe than that in the synclinoria to the south. The continuity of the Dunham and Rugg Brook through both areas, and the extension of the Woods Corners Group slates to the south of the mapped area as an incursion into an otherwise dominantly carbonate basin strongly suggest that this cannot have been the case.

TABLE 2—PARKER SLATE FAUNAS

Base to approximately 450 feet	580 feet and above
Trilobites	Trilobites
Austinvillia billingsi	*Athabaskia* sp.
Bathynotus holopygus	*Chancia rasettii*
Billingsaspis adamsii	*Kochina* sp.
B. walcotti	*Kootenia boucheri*
Bonnia bubaris	*Mexicella stator*
B. capito	*Oriella* sp.
B. senecta (?)	*Peronopsis* sp.
Eoptychoparia kindlei	"*Ptychoparella*" sp.
Kootenia marcoui	*Syspacephalus cadyi*
Olenellus brachycephalus	*Zacanthoides kelsayae*
O. thompsoni	
O. vermontanus	
Onchocephalus? teucer	
Pagetides parkeri	
Periomma typicalis	
P. gaspensis	
Protypus hitchcocki	
Prozacanthoides sp.	
Zacanthopsis sp.	
Brachiopods	Brachiopods
Kutorgina cingulata	"*Finkelnburgia*" sp.
Lingulella granvillensis (?)	*Nisusia festinata*
Nisusia festinata	*Nisusia cloudi*
N. transversa	*Nisusia* (?) *howelli*
Paterina bella	*Undetermined orthoid*
Swantonia antiquata	
Wimanella ? orientalis	
Miscellaneous	Miscellaneous
Anomalocaris ? emmonsi	*Coreospira ? raymondi*
Dactyloidites asteroides	Cystid plates and columnals
D. edsoni	*Hyolithes americanus*
Emmonsaspis cambrensis	sponge spicules
Eocystites? sp.	
Hyolithes americanus (?)	
Planolites congregatus	
P. virgatus	
Protocaris marshi	
Salterella pulchella (?)	
Tuzoia vermontensis	

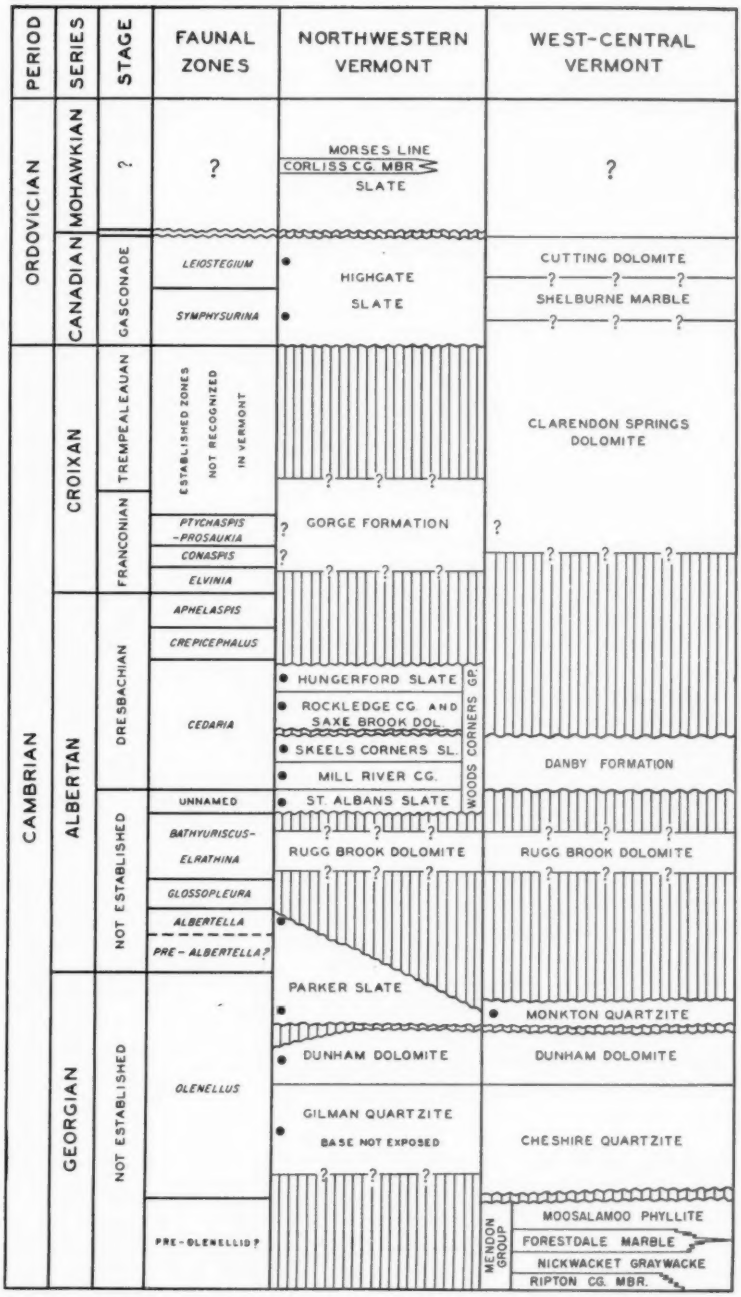

FIGURE 5.—CORRELATION OF THE FORMATIONS OF THE ST. ALBANS AREA AND WEST-CENTRAL VERMONT

Solid circles represent presence of faunas certainly identified with the faunal zone listed to the left. Question marks in Gorge Formation and Clarendon Springs Dolomite represent the *Hungaia magnifica* fauna which seems, at least in part, to correlate with the *Ptychaspis-Prosaukia* and *Conaspis* zones, although younger (and possibly older) rocks are probably present in these units.

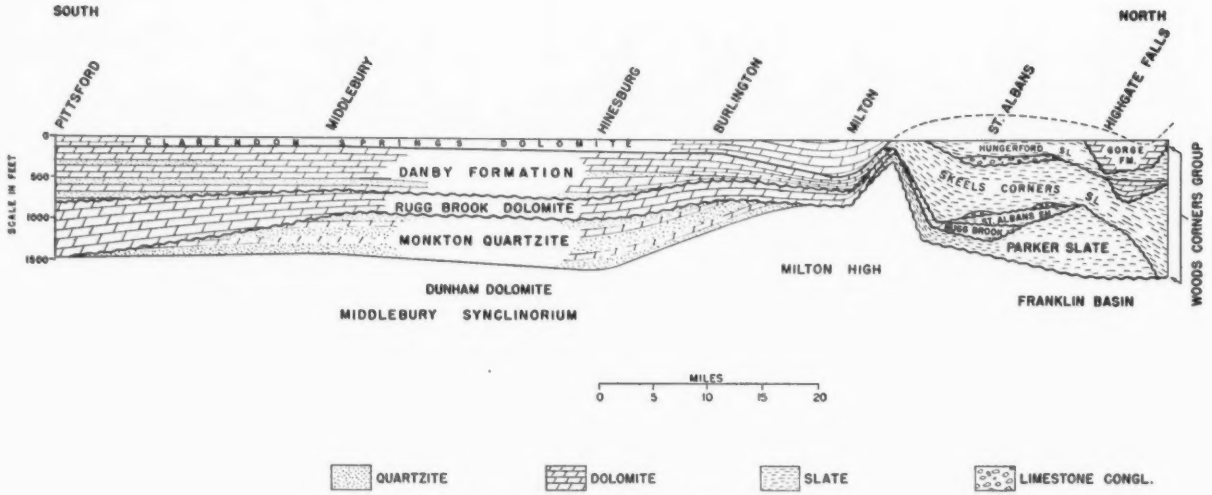

FIGURE 6.—STRATIGRAPHIC DIAGRAM FOR WESTERN VERMONT

Gilman Quartzite and Dunham Dolomite have been omitted. The erosion surface below the Rugg Brook Dolomite is extrapolated south. The erosion surface below the Danby is extended south of Milton on the basis of an erosion surface below the Woods Corners Group. The diagram is roughly parallel to the regional strike of the formations.

EVIDENCE OF THE EXISTENCE OF THE MILTON HIGH: The top of the Cambrian was used as a datum, as nearly as it could be determined (some of the Clarendon Springs Dolomite seems to be Ordovician, but there is insufficient fossil evidence to place the boundary precisely), in the design of Figure 6. The thicknesses in the Franklin Basin were made during this study. From Milton south they are based on data given by Cady (1945, including Fig. 4 and trigonometric estimates from Pl. 10) and previous authors.

The striking fact that appears from this plotting is that between the southern part of the St. Albans area and Milton there is a marked thinning of all formations above the Dunham. The erratic thickening and thinning of the Dunham above the Champlain thrust makes it impossible to decide whether the Dunham thinned over the Milton High; hence the Dunham is omitted from Figure 6. The following should be noted:

(1) The Parker Slate thins southward until only the basal beds carrying a Lower Cambrian fauna extend south of the St. Albans area (Shaw, 1954, Fig. 4; this paper Fig. 3, cols. 1–5). This is due to pre-Rugg Brook (Middle Cambrian) erosion.

(2) The Monkton Formation northeast of Malletts Bay, 11 miles south of the St. Albans area, is 300 feet thick (Keith, quoted by Cady, 1945, p. 531), and Kindle and Tasch (1948) and Tasch (1949) have shown that it is all Lower Cambrian. Shaw (1954, p. 1044) has pointed out that the lower fauna of the Monkton is equivalent to the lowest fauna of the Parker, and an unpublished reexamination of the Monkton collections has shown that the closest similarities are with the collections from the lowest 50 feet of the Parker. About 8 miles south of the St. Albans area the Monkton disappears, and the Rugg Brook Dolomite lies directly on the Dunham. The identity of faunas in the Monkton and Parker indicates that they were once connected, and their absence west of Milton is interpreted here as the result of exposure and removal by erosion during the first movement of the Milton High.

(3) The Rugg Brook Dolomite thins over the Milton High, although the minimum lies at the south edge of the St. Albans area, about 5 miles north of the center of the Monkton-Parker gap.

(4) The Danby-Woods Corners complex also thins over the High, although the minimum is here about halfway between the two previous minima. It is probably significant that the Danby sands and carbonates change facies just at this place into the Woods Corners slates.

Interpretation of the later history of the High will require greater knowledge of the faunas of the Clarendon Springs Dolomite and the Gorge Formation.

In summary, all evidence points to several upward pulsations of the area around Milton, Vermont, during Cambrian time. This shallowing of the sea, or even emergence had a profound effect on the depositional history of the basins to the north and south; the northern basin received the mainly clastic debris and the southern, carbonates. Lochman (1956, p. 1365 ff.) has postulated another homologous basin for formation of the mainly clastic Taconic sequence.

MIDDLE CAMBRIAN

Introduction

The Middle Cambrian is well developed in northwestern Vermont. It includes approximately the upper half of the Parker Slate, the Rugg Brook Dolomite, the St. Albans Slate, and, under the correlation of the *Cedaria* zone adopted herein, the Mill River Conglomerate, Skeels Corners Slate, Rockledge Conglomerate, and the Hungerford Slate. The series attains a maximum aggregate thickness of about 3500 feet and contains the thickest section yet known of rocks belonging to the *Cedaria* zone.

Rugg Brook Dolomite

NAME: The name Rugg Brook Dolomite Conglomerate was proposed by Schuchert (1933, p. 366) for a dolomite and dolomite conglomerate previously confused with the Upper

Cambrian Milton Dolomite, and lying between the Parker Slate and the St. Albans Slate. The most thorough discussion of the formation is that by Howell (1939c), whose interpretations have not been essentially modified by this study.

DISTRIBUTION: The Rugg Brook Dolomite is exposed discontinuously on the west side of the area, and sandy dolomites that are present at a few places along the eastern side of the area (Pl. 1 and Fig. 16) are assigned to the Rugg Brook because of their stratigraphic position and lithology. The Rugg Brook has not been recognized north of the Missisquoi River on the west (Fig. 3), but it extends southward into the Milton quadrangle, where it has been called the Winooski Dolomite by Cady (1945, p. 532).

A thick dolomite conglomerate that forms the east bank of Rugg Brook, about 3 miles southwest of St. Albans, was first chosen as the type locality by Schuchert (1933, p. 366), but this conglomerate proved to be part of the Dunham Dolomite; therefore, Schuchert (1937, p. 1043) chose as the new type locality an exposure of a 10-foot dolomite a thousand feet to the east. However, the best exposures of the Rugg Brook are on the Conner farm, 1.3 miles S. 70° W. of the center of St. Albans. Here, the formation reaches a maximum thickness of about 300 feet as a filling in the channeled top of the Parker Slate.

Booth (1950) has mapped the Rugg Brook on the eastern side of the area, but different interpretations based on a walking-out of the beds are presented on Plate 1 and Figures 12 and 16 of this paper. In the klippe at Bridgeman Hill Booth identified a dolomite within the Skeels Corners Slate as Rugg Brook, and he (1950, p. 1152; Pl. 2, fig. 3) has also identified limestone conglomerate on the northwest slope of Bridgeman Hill as Rugg Brook. Figure 12 is a large-scale map of the area showing the present interpretation. Recognition that the conglomerates are part of the Rockledge Conglomerate removes the difficulties of accounting for this exotic lithology in the Rugg Brook. Likewise, the absence of the Rugg Brook in this area agrees with the regional pattern of distribution.

The thinness of Parker and Skeels Corners slates beneath the Rockledge are accounted for by erosion (as will be described), and by analogy the two patches of limestone conglomerate south of Greens Corners (Pl. 1) seem much better correlated with Rockledge Conglomerate than with Rugg Brook Dolomite (Booth, 1950, Pl. 1).

LITHOLOGY: Howell (1939c, p. 99) has described the Rugg Brook as a "sandy, sometimes conglomeratic, salmon brown to buff brown weathering gray to buff-colored dolomite". Gray arenaceous dolomite comprises most of the rock. Conglomerates containing subangular boulders as much as 2 feet across are locally prominent, as, for example, north of Lake Street, the road between St. Albans and St. Albans Bay; here, both boulders and matrix are of arenaceous dolomite. Some of the dolomites near the northern end of the large outcrop west of St. Albans are faintly pink, a coloring not observed elsewhere in post-Dunham dolomite.

ORIGIN OF CAMBRIAN DOLOMITE CONGLOMERATES: Schuchert (1933, p. 356; 1937, p. 1039) regarded the Rugg Brook as the basal conglomerate of the St. Albans slate, but Howell (1939c, p. 98) has suggested that there is a break between the formations, which this study has tentatively confirmed. Schuchert (1933, p. 366) thought that the boulders came from Dunham or Parker dolomite, but this investigation supports Howell's view (1939c, p. 100) that the conglomerate was derived principally from the Rugg Brook itself and not from older formations; the absence of slate phenoclasts and the uniformly dolomitic nature of the formation, irrespective of the subjacent lithology, cast doubt on the Rugg Brook as a basal conglomerate.

Conglomerates of large blocks of dolomite (almost invariably arenaceous) in an arenaceous dolomite matrix are present in Dunham Dolomite, Parker Slate (below the middle dolomite), Rugg Brook Dolomite, St. Albans Slate and Skeels Corners Slate. All these are here assumed to have had essentially identical modes of origin, *i.e.*, storm action that tore up the blocks from the sea floor. It seems probable that the dolomites were formed in the mud-bottomed lagoons when local shoaling took place. This shallowing first permitted carbonate deposition, then continued after diagenesis, allowing more

vigorous wave action on the sea bottom to form the breccia.

THICKNESS: The Rugg Brook has a maximum thickness of about 300 feet in the Conner Farm area, but elsewhere it is commonly less than 100 feet thick (Fig. 3, cols. 1–5).

STRATIGRAPHIC RELATIONS: Rugg Brook Dolomite lies on the eroded surface of Parker Slate and is in turn overlain by St. Albans Slate. Southward, in the Milton quadrangle (Cady, 1945, Pl. 10) the Rugg Brook has been mapped as Winooski Dolomite.

AGE AND CORRELATION: No fossils are known from the Rugg Brook, but it is Middle Cambrian because it overlies Parker Slate, the upper part of which is of *Albertella* age, and underlies St. Albans Slate, which is apparently of earliest *Cedaria* age or slightly older. The Winooski Dolomite of west-central Vermont is regarded as a southward continuation of the Rugg Brook (as noted by Cady, 1945, p. 522, 533) for the following reasons:

(1) Winooski Dolomite lies on the same post-Parker erosion surface as does Rugg Brook Dolomite in the northern part of the Milton quadrangle and, hence, is Middle Cambrian (Shaw, 1954, p. 1044).
(2) Both formations are essentially alike lithologically, although the Rugg Brook contains more sand and conglomerate than the Winooski (Cady, 1945, p. 533).
(3) The apparent conformity of the Winooski and the underlying Monkton Quartzite can be explained by reworking of the upper beds of the exposed Monkton at the time of the return of the sea in post-*Albertella* time.

In a discussion of the correlation of Rugg Brook and Winooski dolomites with Dr. Cady (August 26, 1954) it was concluded that one name, applied to the whole unit would simplify stratigraphic nomenclature, and that, in view of the complicated and manifold uses of the name Winooski, (reviewed by Cady, 1945, p. 532) the latter name should be dropped and the name Rugg Brook Dolomite be applied in both northwestern and west-central Vermont.

Woods Corners Group

The name Woods Corners Group (taken from the alternate local name for Skeels Corners, 5 miles north of St. Albans) is here proposed for the entire succession between Rugg Brook Dolomite and Gorge Formation. The group, thus, includes St. Albans Slate, Mill River Conglomerate, Skeels Corners Slate, Rockledge Conglomerate and Hungerford Slate. These formations are believed to represent an essentially continuous depositional sequence. It is occasionally necessary to refer to the entire pre-Gorge clastic sequence as a unit, and it is for this purpose that the group name is coined. The name is used in this sense on Figure 6. The name is intended to call attention to the lithic unity of the sequence; the fact that all the formations may be of *Cedaria* age is not a fundamental criterion in the recognition of the group.

St. Albans Slate

NAME: The name St. Albans has been used in many ways, and a thorough review of the variations has been published by Howell (1929). Walcott (1910, p. 255, Figs. 10, 11) first applied the name St. Albans Shale informally to the Middle Cambrian beds near St. Albans, but the name was formally proposed by Howell (1926), who later described the formation in detail (Howell, 1929, p. 265 ff.).

DISTRIBUTION: The type locality is in Adams pasture, at the western edge of St. Albans, north of the road from St. Albans to St. Albans Bay. Unfortunately, the formation is not well exposed here. Elsewhere, the formation has been recognized only on the western side of the area in a narrow belt that disappears northwest of Skeels Corners and does not extend south into the Milton quadrangle (Pl. 1).

LITHOLOGY: St. Albans Slate is commonly black, gray-black or tan and is micaceous. At the type locality the top of the formation is soft, tan to golden-brown micaceous slate, highly cleaved and contorted.

A short distance south of the road leading east from Fonda Quarry, the St. Albans includes a thick conglomerate of large blocks of arenaceous dolomite, averaging 1 x 2 feet on exposed faces, in a shaly matrix, but elsewhere dolomite and dolomite conglomerate are rare. Howell (1929, p. 265) reports thin limestone lenses and nodules, but because of his intensive quarrying

of them none was found during this study. These limestones yielded many of the fossils described by Howell.

Capping the topographic nose 2.55 miles N. 70° E. of the railroad grade crossing at Swanton Junction (Pl. 1) is a mass of limestone conglomerate that is here tentatively assigned to the St. Albans. The lens lies on Parker Slate with slight angular discordance and appears to be stratigraphically above the Rugg Brook, which pinches out a short distance to the south, although the contact relations are not perfectly clear. Slates above the conglomerate have yielded one *Bailiella*, a genus known elsewhere in the St. Albans Slate, proving that the lens cannot belong to the Mill River, with which its appearance obviously associates it. Field relations do not exclude the possibility that the lens is a local development within the Rugg Brook, but the writer does not believe this is the case. The pebbles are well-rounded to subangular blocks of blue limestone, which yield dark blue-black spheroids 1–1½ mm in diameter; these spheroids seem to be organic and are interpreted as coprolites rather than as oolites. The matrix is a tan dolomite. All attempts to find fossils in either the pebbles or the matrix were unsuccessful.

THICKNESS: Schuchert (1937, p. 1041) estimated that there were 270 feet of St. Albans Slate in Adams pasture, but in view of the crumpling of the beds, an estimate of about 200 feet ± 50 per cent seems more realistic. The belt of outcrop widens northward (Pl. 1), but the structure exposed in the bed of Stevens Brook suggests that the beds have flattened out, and that the formation has not thickened in that direction. Elsewhere than at the type locality the formation is commonly less than 100 feet thick (Fig. 3, cols. 2–5).

STRATIGRAPHIC RELATIONS: The contact between St. Albans Slate and Rugg Brook Dolomite, below, is obscure. West of Adams pasture the St. Albans seems to be disconformable on the Rugg Brook (Howell, 1929, p. 266), but because of the disturbed condition of the slates there can be no real certainty on this point. Elsewhere the lower contact is not exposed.

The contact between the St. Albans and the overlying Mill River Conglomerate and Skeels Corners Slate will be discussed on a later page.

FAUNA AND AGE: The fauna of the St. Albans Slate is in a very confused state because collections that have been made in the past have been obtained in part from boulders, in part from pebbles in Mill River Conglomerate, and from exposures now assigned to Skeels Corners Slate, as well as from beds *in situ*. Any attempt to review thoroughly all the species reported by Howell (1937) from the St. Albans would unduly extend this part of the paper, but a preliminary examination of specimens from rocks known to be in place indicates the presence of the following: *Alokistocarella?* n. sp., *Armonia? franklinensis*, *Bailiella* sp. (float), *Centropleura vermontensis*, *Hastagnostus angustus*, *Hypagnostus franklinensis*, *Onchonotopsis eminens*, *?Solenopleura vermontensis*, *?Lingulepis vermontensis*, *Protorthis vermontensis*, *Helcionella vermontensis*.

This fauna and the other species described by Howell (1937) seem to present a mixture of Middle Cambrian types with those commonly found in the basal Dresbachian *Cedaria* zone. This implication that the *Cedaria* zone may, thus, be equivalent to part of the Middle Cambrian of Europe (as already suggested by Wilson, 1954) is discussed on a later page.

Mill River Conglomerate

NAME AND DISTRIBUTION: The name Mill River Conglomerate was proposed by Keith (*in* Howell, 1929, p. 266, footnote 1) for the limestone conglomerate lenses that lie between St. Albans and Skeels Corners slates. The formation is discontinuous and is exposed at only five localities, all on the western side of the area (Pl. 1). The most southerly, a low bluff 0.89 mile S. 65° E. of Everett School, Georgia Township, is the type locality. The remaining exposures are (1) a knoll 1.09 miles N. 83° E. of Everett School; (2) in Adams pasture, 0.95 mile N. 77° W. of the center of St. Albans (intersection of Lake and Main Streets); (3) on the east side of the St. Albans-Swanton road 0.61 mile S. 71° W. of Bench Mark 271 on the railroad bridge over Stevens Brook; and (4) a low ridge 0.97 mile S. 61° E. of the railroad grade crossing at Swanton Junction.

LITHOLOGY: Mill River Conglomerate, as it is

here restricted, contains only limestone conglomerates; it is composed largely of rounded to subangular boulders and pebbles of limestone in a matrix of arenaceous blue-gray limestone and red-brown dolomite. Shale phenoclasts are rare. Howell (1939b) mentions the presence of arenaceous dolomites in the formation, but as here interpreted there are no such beds in the Mill River.

The limestone pebbles are of three types: (1) a pale-blue-gray massive variety derived from bioherms and ranging in size from small rounded pebbles less than 1-inch in greatest dimension to lentils 15 feet long; (2) a blue-black, massive, apparently non-biohermal variety that makes small sub-rounded fragments, which are only exceptionally more than 6 inches long; and (3) a blue-black bedded variety that also forms sub-rounded pebbles, commonly less than 6 inches in length. The biohermal blocks are not fossiliferous, and the bedded pebbles are only sparingly so, but the dark massive pebbles are commonly crowded with trilobites. The limestones must have been indurated before being broken up, because fossils terminate abruptly at the margins of the pebbles but are undamaged within.

ORIGIN OF CONGLOMERATE: Schuchert (1937, p. 1047) regarded the Mill River Conglomerate as having been formed by the destruction of Lower Cambrian limestone lentils by an invading sea—a true "basal breccia". But, as Schuchert did not know, the pebbles are not Lower Cambrian. All fossiliferous pebbles collected during this study contained Dresbachian fossils, and although Howell (1937) has mentioned the presence of older Middle Cambrian pebbles, nothing older has been seen during this investigation .

Mill River Conglomerate has many lithologic features in common with Rockledge Conglomerate, and it is here contended that they were both formed in shallow water around shoals or islands.

Three facts must be borne in mind when considering the origin of Mill River Conglomerate: (1) the absence of shale pebbles in the conglomerate; (2) the presence of large numbers of limestone fragments in the conglomerate, whereas limestone is almost absent in underlying St. Albans Shales; and (3) the presence of considerable quantities of coarse sand in Mill River Conglomerate and basal Skeels Corners Slate and its rarity in the St. Albans.

The following sequence of events may explain these facts. After the initial invasion by the St. Albans sea there was a slight shallowing, without folding. Limestone bioherms formed around the resulting shoals, and debris from them formed clastic, fossiliferous limestone in protected areas. Later, the bioherms and the limestones were both exposed, either by uplift or further shoaling, and broken up to form the conglomerate. Synchronous exposure of the shorelands surrounding the St. Albans sea to the west and north, and probably also to the south and southwest along the Milton High supplied sands, probably in part from the Rugg Brook Dolomite, but undoubtedly also from source areas now no longer to be seen because of of the overthrusting. (The northern or northwestern source areas continued to supply sand during the deposition of Skeels Corners Slate and later during formation of Saxe Brook Dolomite). Renewed transgression and deepening of the sea finally overwhelmed the local limestone areas with the muddy sediments of the Skeels Corners. The general absence of slate fragments in the conglomerate can be explained by the fact that the St. Albans was not exposed for a prolonged period but remained most of the time below sea level, whereas on the shoals the muds were covered by a limestone capping. The slates are believed to be the result of deeper water conditions such as those postulated by Lochman (1956, p. 1348, 1363) for the clastics of the Taconic sequence.

THICKNESS: The Mill River forms discontinuous lenses, commonly less than 200 feet long and no more than 35 feet thick.

STRATIGRAPHIC RELATIONS: The contact between Mill River Conglomerate and St. Albans Slate is poorly exposed, but there seems to be no evidence to suggest that it is not conformable. The conglomerate grades upward into overlying Skeels Corners Slate, and the Mill River might justifiably be regarded as a member of the Skeels Corners, but since the Mill River has received previous recognition as a formation there seems to be little gain in reducing it to a member.

FAUNA AND AGE: The fauna of the pebbles in

the Mill River has been studied by Mary Gilman Clark and the author and will be described in a separate paper. The pebbles contain a fauna typical of the *Cedaria* zone and include, in addition to the undescribed forms, *Blountia* (*Homodictya*), *Blountiella?*, *Bolaspidella*, *Brassicicephalus*, *Bynumia*, *Catillicephala*, *Cryptoderaspis*, *Homagnostus* (or *Proagnostus*), *Kingsstonia?*, *Kormagnostus*, *Onchonotopsis Leiocoryphe*, *Metisaspis*, *Uncaspis*, and *Welleraspis*. The Mill River Conglomerate, like the Skeels Corners Slate above it, was first described as an Upper Cambrian unit, but this correlation is now questioned and will be discussed below.

Skeels Corners Slate

NAME: Skeels Corners Slate includes some of the beds formerly called Georgia Slates by Walcott (1886) and a large part of those referred to Hungerford and Highgate Slates by Keith (1932), Schuchert (1937) and Raymond (1937). The name was originally proposed by Howell (1939a) to refer to the *Bovicornellum* beds alone, but it is here extended to include all strata lying above Mill River Conglomerate (or, in its absence, St. Albans Slate) and below Rockledge Conglomerate. The type locality is half a mile northwest of Skeels Corners and 6 miles north of St. Albans.

DISTRIBUTION: The greatest development of Skeels Corners Slate is in the south-central part of the St. Albans quadrangle, where it underlies the entire center of the area south of St. Albans Hill, in a band 2 miles wide. Northward, exposures are rare except north of a point about 3¼ miles north of St. Albans, and at the type locality.

North of the Missisquoi River the slates above the upper Parker limestone conglomerate and below Saxe Brook Dolomite on the east side of Saxe Brook valley are tentatively assigned to the Skeels Corners. These beds are fossiliferous, for a 2-day search yielded two complete trilobites and three thoraces, but all were badly weathered, Howell (Personal communication) has collected *Bovicornellum* beneath Saxe Brook Dolomite on the east side of Saxe Brook valley, but none could be found during the present survey.

Along the eastern side of the St. Albans area it is more difficult to trace the Skeels Corners accurately because carbonate rocks are not present to separate the formation from underlying Parker Slate or overlying Hungerford Slate; therefore, the boundaries shown on Plate 1 are to be regarded as provisional. The discovery of *Bovicornellum vermontense* in rocks in place southeast of St. Albans Hill, by P. T. Fowler in 1946 proves the existence of the formation of the eastern side of the Rosenberg syncline, however.

LITHOLOGY: The Skeels Corners is typically a black slate, although there are local developments of dolomite, sandstone, dolomite conglomerate, limestone bioherms, limestone, and calcareous shale in minor quantities.

Because of the local variation in development of the minor constituents it seems best to describe four areas of outcrop separately; these are: (1) southwest of St. Albans Hill, (2) the type locality north of Skeels Corners, (3) beds assigned to the formation in the Saxe Brook valley, and (4) the slates along the eastern side of the area.

Southwest of St. Albans Hill, the lower 400–600 feet of the formation is composed of arenaceous, slightly calcareous gray shale and slate (Fig. 3, cols. 1, 2). Approximately 200 feet above the base of the formation there is a zone of blue limestone bioherms (Schuchert, 1937, Pl. 5, fig. 1) some of which are fossiliferous. The zone contains nine bioherms and a small amount of dark-blue bedded limestone. The upper slates have little quartz sand in them and no limestone save one cluster of three small bioherms. In both the arenaceous and non-arenaceous slates the beds are commonly 5–10 mm thick and are separated by white sand stone laminae less than 1 mm thick; the two alternate in endless succession. The thin beds of limonite-stained sandstone so common in the section west of Georgia Center, 3 miles to the south in the Milton quadrangle, are lacking or very scarce. A limestone conglomerate crops out in the lower part of the formation 300 yards N. 30° E. of the type locality of the Mill River Conglomerate, which is 0.89 mile S. 65° E. of Everett School. It is composed of subangular limestone blocks 1–10 inches across in a gray shale matrix, and it closely resembles Mill River Conglomerate, but the matrix contains

Bovicornellum vermontense, the guide fossil for Skeels Corners Slate.

Black slate is exposed in the east-central ninth of the St. Albans quadrangle and includes a great many thick lenses of sandy dolomite and dolomite conglomerate. There are a few limestone bioherms, and one of these is the huge mass that gives the Rockledge Estate its name. This bioherm has never been part of the Rockledge Conglomerate as Schuchert (1937, p. 1064) implied.

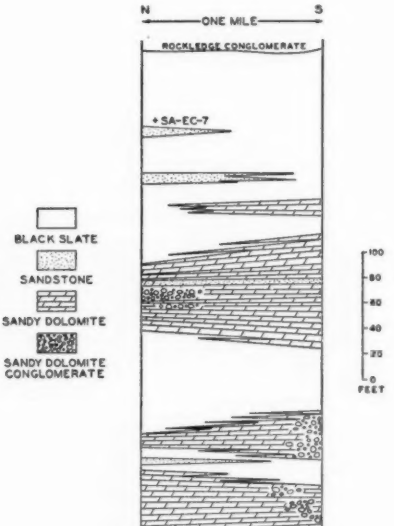

FIGURE 7.—GENERALIZED COLUMNAR SECTION OF SKEELS CORNERS SLATE AT TYPE LOCALITY

Half a mile northwest of Skeels Corners. The cross marked SA-EC-7 is a fossil locality 375 feet above the base of the formation. Its fauna is listed in the tenth column of Table 3.

At the type locality, northwest of Skeels Corners, there is an exceptional local development of sandstone, sandy dolomite, dolomite conglomerate, and dolomite, all of which are relatively uncommon elsewhere in the formation (Figs. 7, 9). The conglomerate consists of a matrix of arenaceous, brown-weathering, gray dolomite containing blocks of similar arenaceous dolomite as much as 2 feet across, but which average 1 foot in greatest dimension. The upper quartzites seem to be replaced southward by shales since they do not crop out in that direction. The shales in the area covered by Figures 7 and 9 are mainly black or gray and in part dolomitic. The uppermost slates, are mainly gray and in part micaceous, but the fossiliferous beds are black and contain many blebs and pockets of red iron oxide. Bedding is irregular in the fossiliferous layers but is generally regular in the other strata.

The rocks referred to Skeels Corners Slate in the Saxe Brook valley (Pl. 1) are predominantly well-bedded gray shales, containing dolomite and quartzite beds up to an inch thick. These beds may not be properly assignable to the Skeels Corners, in whole or in part. Thin-bedded black slate crops out on the west side of Carter Hill and intermittently for about a mile to the north, and Dr. Howell has written the author (July 23, 1947) that he found *Bovicornellum* in them "about one and a quarter miles north-northeast of the 'C' in 'Carter Hill' as that name is printed on the St. Albans Quadrangle topographic map". Although an intensive search was made in this area no additional fossils were found during this study.

On the eastern side of the St. Albans area the beds assigned to the formation are black, thin-bedded slate containing thin quartzites, which in many places make up about half of the rock. These slates are commonly thrown into drag folds. As has been mentioned above, it is usually difficult in the east to separate the Skeels Corners Slates from those in the Parker and Hungerford. In general, however, Skeels Corners Slate is less dolomitic than underlying Parker Slate, which weathers to more rounded surfaces. The Parker is blacker, more intensely cleaved and lacks the calcareous beds of the rocks referred to the Hungerford. It must be emphasized that these differences are not infallible guides to these formations, and most of the outcrops are recognized by the stratigraphic order or by lateral tracing.

THICKNESS: It is impossible to obtain accurate measurements of the thickness of Skeels Corners Slate, and the thicknesses shown on Figure 3 are to be regarded only as likely estimates; an error in some of 50 per cent is not impossible. The formation is thickest in the southern part of the St. Albans quadrangle and

the northern part of the adjoining Milton quadrangle, where it reaches an estimated 1200 feet. It thins both northward and eastward, due in part to later erosion and possibly also to original thinning.

Where Mill River Conglomerate is absent the Skeels Corners lies on St. Albans Slate or an older formation. Inasmuch as a slight disconformity is assumed to be at the base of the Mill River there is presumably an identical surface

TABLE 3.—PRELIMINARY LIST OF FOSSILS IN THE SKEELS CORNERS SLATE

Numbers in the boxes indicate the number of specimens. C = Cranidia. CK = Free cheek. P = Pygidium. E = Entire specimens. M = Meraspides. N = Numerous. Note that the use of the names *Catillicephala rotunda* and *Hemirhodon viator* indicates forms present, but does not imply acceptance or rejection of the validity of these species.

SPECIES \ Elevations above base of formation	35'	35'	72'	80'	100'±	130'	155'	170'	215'	375'	470'	810'
Bolaspidella macgerriglei (Raymond), 1937				2C 1Ck				1E, 7C 2T, 2M				
Bowicomellum vermontense Howell, 1934	20	1	2		5	1	1	25+	N	105	8	
Bynumia metisensis Rasetti, 1946										1C		
Catillicephala lata (Raymond), 1937										6C		
C rotunda (Rasetti), 1946										4C 1P		
C. sp. indet.	1P			1P						17C 12P	1P	
Hemirhodon schucherti Raymond, 1937												4C 1P
H. viator Raymond, 1937												3C
H. sp. indet.												12P
Lingulella? stonei Shaw, 1953						18	23	1				
Lingulella vermontensis Howell, 1937				21								
Solenopleura? franklinensis Howell, 1937				1C								
Solenopleura? vermontensis Howell, 1937				1C 3P, 2Ch						2C		
Acheilus sp.										1C		
Hyolithes sp.								1				
Hypseloconus? sp.										2		
Oboloid brachiopods						6	3			9		
Cystid plates										21		
Saukiinid pygidium										1		
Unassigned ptychoparid pygidia	1									1		

STRATIGRAPHIC RELATIONS: The contact between Skeels Corners Slate and Mill River Conglomerate is gradational. This may be seen in all outcrops of the Mill River except that which lies about 1½ miles north of St. Albans. Upward from the Mill River the sandy limestone matrix is replaced by gray shale. The pebbles in this shale are commonly about 2 inches across and sub-rounded to well-rounded. A few small well-worn pebbles may be found isolated in the Skeels Corners as much as 10 feet above the main mass of the conglomerate.

at the base of the Skeels Corners where the Mill River is absent, but no direct field observations have been made to support or deny this assumption. The Skeels Corners is regionally unconformable upon the older formations.

FAUNA AND AGE: The fossils of the Skeels Corners Slate have heretofore been referred to the Hungerford or Highgate slates because of the confusion that has existed about the stratigraphic position of those units. Table 3 gives a preliminary determination of the Skeels Corners fauna, of which the most common

member is the little horned-shaped tube, *Bovicornellum vermontense*. This fossil is present in at least the lower 500 feet of the formation. The presence of such genera as *Bynumia*, *Catillicephala*, and *Hemirhodon* as well as the presence of *Cedaria* zone fossils in underlying Mill River Conglomerate indicate clearly that the Skeels Corners is assignable to the *Cedaria* zone, which is here assigned to the Middle Cambrian.

Rockledge Conglomerate

NAME: The name Rockledge Limestone Breccia was proposed by Schuchert (1937, p. 1049) when it was realized that the names Swanton Conglomerate (Keith, 1923, p. 118) and Corliss Conglomerate (Keith, 1932, p. 378; Schuchert, 1933, p. 378–379) had been applied to limestone conglomerates at several stratigraphic levels. Schuchert applied the name Rockledge to the presumed Upper Cambrian rocks and restricted the Corliss to certain Ordovician conglomerates. The Swanton was dropped. Schuchert's longer name has generally been shortened to the simpler phrase Rockledge Conglomerate.

DISTRIBUTION: The type locality is on Rockledge Estate 4.4 miles north of the center of St. Albans, on the west side of the road to Highgate Center. It should be noted that the large limestone bioherm from which the Estate takes its name is not in Rockledge Conglomerate but in the top of underlying Skeels Corners Slate. Northward from the type locality the formation extends in unbroken outcrop to a point about 1¾ miles S. 20° W. of Highgate Falls. Southward, the Rockledge crops out in discontinuous lenses. The most southerly exposure in the St. Albans quadrangle is on the west and northwest flanks of St. Albans Hill, but there is extensive development of the formation in the Milton quadrangle.

Two small patches of limestone conglomerate 0.8 mile S. 10° W. of Greens Corners, 4 miles northeast of St. Albans, and another larger outcrop on the northeast slope of Bridgeman Hill (Fig. 12) seem to occupy the appropriate stratigraphic position to be correlated with the Rockledge. Booth (1950, p. 1152, Pl. 2, fig. 3) has placed these conglomerates in Rugg Brook Dolomite, but as was discussed above, they do not show Rugg Brook lithology, and they fit more simply into the stratigraphic picture if they are regarded as Rockledge. Cady (Personal communication, August 26, 1954) agrees that these conglomerates are Rockledge.

LITHOLOGY: Schuchert (1937, p. 1015, 1050-1052, 1062, 1063–1064, Pl. 1, Pl. 6, fig. 1) has given an extensive description of the Rockledge in most of its outcrops, and although the present study has led to different structural interpretations from those that Schuchert set forth and to a different hypothesis for the origin of the conglomerate, his description makes similar treatment unnecessary here. In most outcrops the Rockledge has at its base a bed of highly arenaceous massive gray dolomite which locally changes to an indurated gray sandstone. This bed is persistent and is present in most exposures from the type locality south to the Milton quadrangle. It is of irregular thickness but increases northward from an average of about 4 feet in the northern part of the Milton quadrangle to about 50 feet north of Skeels Corners (Fig. 3, col. 5). The basal sandy layer grades upward into the conglomerate, which, in turn, has a matrix of dark-gray limestone with much well-rounded frosted quartz sand and averages about 0.5 mm in diameter. The phenoclasts are principally fragments of biohermal limestone ranging in size from small pebbles to huge masses nearly 100 feet long. Far less common than the biohermal fragments, but still abundant, are slabby pebbles of dark-gray bedded limestone 1–4 inches in diameter, and, finally, there are irregular pebbles of dark-gray massive arenaceous limestone; the last two types contain most of the fossils recovered from pebbles; the latter is more productive. One dolomite boulder about 4 feet square was seen, but in general the conglomerate is notable for its lack of dolomite fragments.

The exposures on St. Albans Hill and three lenses in the eastern belt of outcrop are of basically similar rock, but they have been subjected to such intense shearing that the pebbles are stretched, and those on St. Albans Hill have been pulled out into thin sheets that resemble bedding.

Rockledge Conglomerate is lithologically indistinguishable from Mill River Conglomerate. Both have similar biohermal limestone blocks, bedded and massive limestones, and a very sandy matrix. The pebbles contain different

faunas, however. Both the Rockledge and Mill River differ from younger conglomerates, which normally contain little sand, have a smaller proportion of matrix and in many cases lack biohermal fragments.

THICKNESS: Because of the lensing of the Rockledge, no generally applicable figure can be given for thickness (Fig. 3, cols. 2–5). The maximum thickness in the St. Albans quadrangle is south of Hungerford Brook, where it is about 100 feet, of which the lower sandy dolomites comprise about half. In most outcrops, however, the formation is only 30–50 feet thick.

STRATIGRAPHIC RELATIONS: A discussion of the stratigraphic relations of the Rockledge is postponed until Saxe Brook Dolomite has been described because the two seem to be facies equivalents.

FAUNA: The fauna of the Rockledge Conglomerate has been treated in detail in a recent article (Shaw, 1952) and is summarized in Table 4. Both the matrix and pebbles of the Rockledge seem to be best correlated with the *Cedaria* zone rather than the *Crepicephalus* zone as suggested by Lochman and Duncan (1944, p. 35). Earlier reports (Keith, 1923, p. 119) cited the presence of Lower Cambrian pebbles in the Rockledge, but Schuchert (1937, p. 1014) stated that this was incorrect, and the present study has yielded only Dresbachian forms.

Saxe Brook Dolomite

NAME AND DISTRIBUTION: The name Saxe Brook was proposed by Howell (1939a) to include several dolomites that overlie the *Bovicornellum* beds. The dolomites south of the Missisquoi River that were originally included in the formation are now placed in the amended Skeels Corners Slate, and Saxe Brook Dolomite is limited to the beds that form Carter Hill, a mile west of Highgate Center, and the ridge to the north, on the east side of Saxe Brook valley (Pl. 1.). The Saxe Brook has not been recognized on the east side of the area, and it extends into Canada only in a much reduced thickness (Fig. 3, cols. 7–11). Saxe Brook Dolomite is here regarded as the northern equivalent of Rockledge Conglomerate, but the separate names are retained because of the lithologic differences between the two units.

LITHOLOGY: The Saxe Brook contains three main rock types. The most common is a highly arenaceous gray dolomite, the next is a sandstone with dolomitic cement, and the least common is non-arenaceous dolomite. These types are arbitrary divisions of a continuous, gradational lithic sequence. The distribution of the three types is shown diagrammatically in Figure 8. The sandstone and sandy dolomite are well bedded but the sandless dolomite is not. The weathered surfaces of all arenaceous rocks tend to develop a leached "crust" of poorly cemented sand grains, which must be broken away to reveal the fresh rock beneath.

TABLE 4.—FAUNA OF THE ROCKLEDGE CONGLOMERATE

Slightly modified after Shaw, 1952

Species	Matrix	Pebbles
Blountia (Homodictya) imitatrix (= *Acrohybus argutus*)	X	X
Catillicephala lata	X	X
C. fowleri	X	X
Coosia? sp.	X	X
Kingstonia scrinium	X	X
Leiocoryphe? prisca?	X	X
Meteoraspis? minuta	X	X
Agnostid No. 1	X	X
Acmarhachis sp.	X	
Bolaspidella macgerriglei	X	
Distazeris simplex	X	
Vermontella? minor	X	
Welleraspis sp.	X	
Pygidium No. 2	X	
Pygidium No. 5	X	
Pygidium No. 6	X	
Bynumia leptogaster		X
"*Hypseloconus*" sp.		X
Maryvillia arion (= *Protillaenus marginatus*)		X
Stenelymus kobayashii		X
Theodenisia, n.sp.		X
Pygidium No. 3		X
Free Cheek No. 1		X

ORIGIN OF SAXE BROOK DOLOMITE: Figure 8 shows a concentration of sand in the southern part of the Saxe Brook, which implies a source of sand to the south or southwest. Similarly, the presence of sand to the north indicates another source to the north or northwest. The possibility of an eastern source is precluded by the

absence of the formation in that direction, although Rockledge Conglomerate crops out sparsely there. It is possible that all the sand came from a single source to the west, but the existence of two separate sandy zones in the Saxe Brook, separated by a relatively sand-free interval more strongly suggests separate sources.

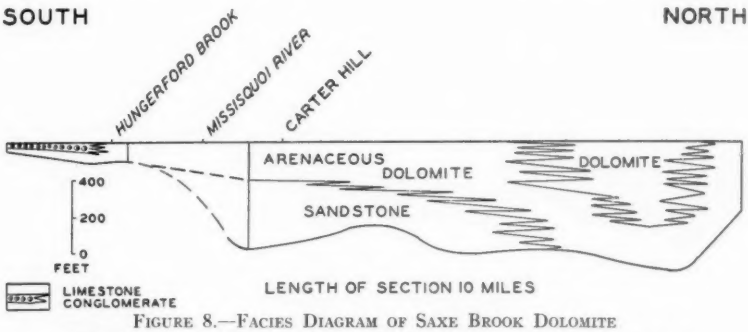

FIGURE 8.—FACIES DIAGRAM OF SAXE BROOK DOLOMITE
Showing assumed relations to Rockledge conglomerate. *See* discussion in text.

The Saxe Brook may, in a sense, be regarded as an arrested stage in Rockledge development —a sort of overgrown basal clastic unit. The sea floor north of the present Missisquoi River probably rose in response to the same stresses that caused the warping of the Rockledge lagoon, but the elevation was apparently never enough to permit the formation of the bioherms and conglomerate found to the south. The southern Saxe Brook sandstones could have been derived in part from the erosion of the thick quartzites at the type locality of Skeels Corners Slate, but probably a western or southwestern source supplied most of the clastics.

THICKNESS: Saxe Brook Dolomite is of rather constant thickness although the outcrop pattern widens west of Cutler Pond because of a fault (Pl. 1). At Carter Hill, 1.3 miles N. 30° W. of Highgate Center, the formation is about 600 feet thick; thence it gradually increases northward to about 700 feet at 45°00′ N. Lat. and then thins abruptly to about 350 feet at the International Boundary (Fig. 8; Fig. 3, cols. 7–11).

STRATIGRAPHIC RELATIONS: Rockledge Conglomerate and Saxe Brook Dolomite both lie above Skeels Corners Slate with slight angular unconformity. Schuchert (1937, p. 1060, 1064, 1065, *etc.*) noted the existence of this unconformity, but in his summary table and general discussion of the Upper Cambrian he (1937, p. 1021, 1050) referred to the contact as an "erosional unconformity" or a "disconformity". The most southerly exposure of this contact in the St. Albans area is 1.4 miles S. 50° E. of the railroad grade crossing at Swanton Junction, where Skeels Corners Slate is overlain by shale beds at the base of the Rockledge Conglomerate. The Rockledge cuts out about 6 feet of Skeels Corners; also the Rockledge dips more steeply eastward, showing that the Skeels Corners had a slight westerly dip when the Rockledge was deposited. The unconformity is also well shown at the type locality of Skeels Corners Slate, half a mile northwest of Skeels Corners (Fig. 9). Here, the Skeels Corners is warped into a local anticline and syncline whose axes (particularly the synclinal axis) do not correspond with those of folds in the overlying Rockledge Conglomerate and clearly demonstrate the presence of folds in the Skeels Corners before the deposition of the Rockledge.

A third exposure of the unconformity lies below Saxe Brook Dolomite in Saxe Brook valley; here, in a strip more than half a mile long, from 2 to 2½ miles N. 20° W. of Highgate Center, the unconformable contact is exposed at a number of places. The angular discordance in strike between the slate and the overlying dolomite is about 50°; dips are similar or are steeper in the slates. It is possible that the angularity of this contact has been increased by northward movement of the Saxe Brook during folding, but there can be no doubt of the origi-

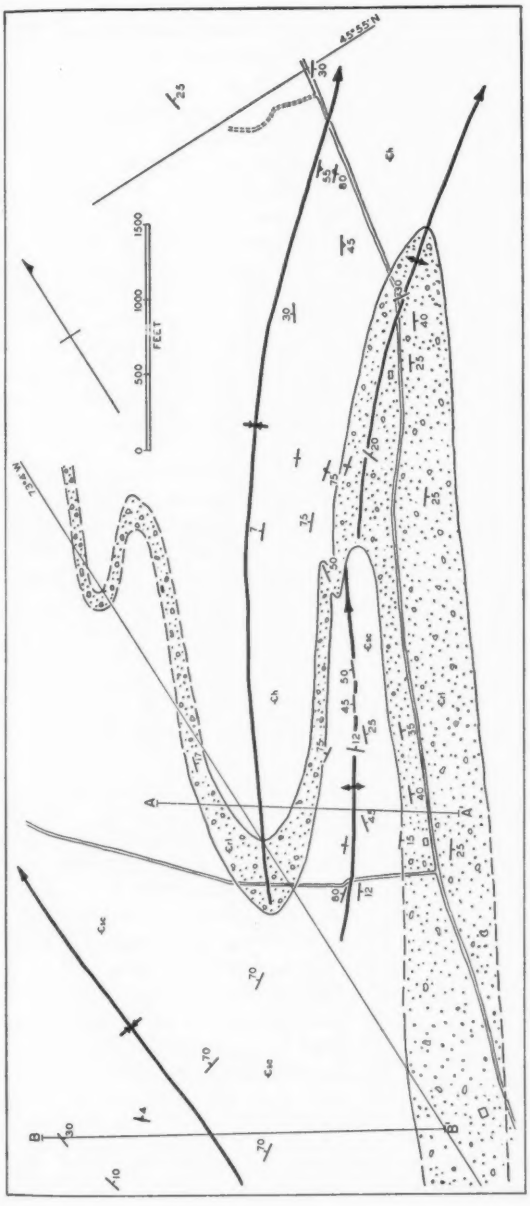

FIGURE 9.—GEOLOGICAL MAP AND SECTIONS OF THE HUNGERFORD BROOK-DONALDSON LANE AREA

The northern two-thirds of this map is a modification of a plane-table map made in 1942 under the supervision of Dr. M. P. Billings and the late Dr. P. E. Raymond. €sc = Skeels Corners Slate. €rl = Rockledge Conglomerate. €h = Hungerford Slate.

nal discordance of the two formations. The movements of the Saxe Brook will be discussed more fully as part of the description of the Saxe Brook tear fault.

FAUNA: No fossils have been collected from the Saxe Brook Dolomite, but is assumed to be Dresbachian on the grounds that it is equivalent to the Rockledge Conglomerate.

Origin of Rockledge Conglomerate

Theories to explain the origin of Rockledge Conglomerate have been expounded by Schuchert (1937), Raymond (1937), and Cady (1945), but all are somewhat different from that proposed here. Several facts must be accounted for in any explanation of the origin of the Rockledge: (1) the angular unconformity beneath the Rockledge; (2) the basal sandy dolomite; (3) the non-biohermal limestone pebbles; (4) the presence above the basal clastics of a discontinuous zone containing bioherms that appear to be in place; (5) sand in the matrix of the conglomerate; (6) the absence of shale and dolomite phenoclasts; (7) the patchy distribution of the conglomerate; (8) the gradational contact with the Hungerford; and (9) the similarity of the faunas in the pebbles, matrix, overlying and underlying beds.

All these facts can be explained if we assume that the following events occurred:

(1) Slight warping took place at the end of Skeels Corners time, and shoals or low islands were raised from the sea bottom. This accounts for the angular unconformity beneath the Rockledge.

(2) As these rising shoals reached wave base the sand in Skeels Corners Slate was reworked, and the finer argillaceous material was winnowed out and carried away to be deposited in the unelevated parts of the sea floor. The shoaling was accompanied by deposition of carbonate that mixed with the reworked sand to form the basal clastic rocks now found in almost all outcrops. External sand sources were also probably operative both north and south.

(3) Bioherms developed and formed protected areas in which bedded fossiliferous limestone was deposited; the biohermal limestone covered the older beds and prevented contamination of the bedded limestone by sand.

(4) Further shoaling raised the bioherms sufficiently above wave base to permit wave action to destroy them and the bedded limestone they protected, thus forming the conglomerate. Destruction was not complete, however, and remains of the bioherm zone were left just above the basal clastic and below the main body of the conglomerate.

(5) Destruction of the bioherms and the accompanying limestone re-exposed at least part of the basal clastics of the Rockledge and possibly the sands of the Skeels Corners and other sources north and south which supplied sand to the matrix of the developing conglomerate.

(6) Shoaling ceased, and a slight rise in relative sea level or a sinking of the sea bottom permitted conditions of muddy sedimentation which had persisted on the intershoal lagoonal bottom and in the geosynclinal areas to the east to spread over the shoals again and bury the conglomerate.

This hypothesis also explains the patchy distribution of the Rockledge. The conglomerate would develop only where shoaling permitted the growth of bioherms and the formation of limestone. At the margin of a shoal there might be deposition of only the basal clastic layer; this actually seems to have occurred north of the Skeels Corners area, where the basal beds are thick and farther north, across the Missisquoi River, where these beds appear as the sandstone and arenaceous dolomite of Saxe Brook Dolomite (Fig. 8).

The gradational contact between the Rockledge and overlying Hungerford Slate is the obvious consequence of the end of shoaling; the Hungerford represents the re-establishment of the muddy conditions that prevailed during Skeels Corners time. Therefore, the unconformity at the base of the Rockledge in the midst of beds that seem elsewhere to represent essentially continuous deposition is unusual. The less common occurrence of the Rockledge eastward is likely due to the gradual deepening of the shelf toward the geosyncline and correspondingly fewer shoals. The assumed continuity of deposition in the non-shoal areas accounts for the difficulty encountered in trying to distinguish between Skeels Corners and Hungerford slates on purely lithologic grounds; according to this hypothesis there was no fundamental difference between them originally.

Hungerford Slate

NAME: The name Hungerford Slate was proposed by Schuchert (1937, p. 1047–1048) when he recognized that part of the slates previously referred to the Lower Ordovician Highgate Formation were actually Upper Cambrian. Schuchert (1037, p. 1948) states that, "The name is taken from Hungerford Brook, 1.25 miles south of Highgate, the type area being where the brook crosses the Highgate road . . ." Schuchert misinterpreted the structures at the type locality, however, and believed that the Hungerford Slate lay below the Rockledge Conglomerate.

In 1942, this area was mapped by plane-table under the direction of M. P. Billings and the late P. E. Raymond, who have kindly permitted the use of the resulting map. This map has been incorporated as the northern three-quarters of Figure 9 of this report; the detailed mapping proved that the Rockledge plunges northward under the slate at Hungerford Brook and that Hungeriord Slate is, therefore, younger than the Rockledge. Because of the misinterpretation of the stratigraphic position of the Hungerford Slate, Schuchert and subsequent workers have referred to the Hungerford Slate all the beds now designated as Skeels Corners Slate.

Ruedemann (1947, p. 165), in describing the graptolite *Dictyonema schucherti*, referred to this formation as Russell Slate, but the name is not necessary and is probably not valid, for it was not defined.

DISTRIBUTION: The distribution of Hungerford Slate is not well known because large areas that are thought to be underlain by it are covered with glacial debris. Moreover, erosion in Franconian and again in Trempealeauan time cut deeply into its upper beds, and the formation is also partly overridden by other rocks in the Gore and Highgate slices (Fig. 2). In view of these complications the upper formational boundaries shown on the areal map (Pl. 1) and the thicknesses shown on Figure 3 are in some places no more than approximations.

The narrow valley containing Cutler Pond, 3 miles north of Highgate Center, is floored by Hungerford Slate. The narrowing of this valley southward is a topographic expression of an erosional unconformity on top of the Hungerford.

South of the Missisquoi River, along the Highgate-St. Albans road, from Hungerford Brook south to St. Albans Hill, the base of the Hungerford is exposed above Rockledge Conglomerate.

On the eastern side of the area well-bedded slates are referred to the Hungerford because of their position above limestone conglomerates assigned to the Rockledge.

LITHOLOGY: The Hungerford consists almost entirely of black slate, most commonly in beds 2–10 mm thick, alternating with beds of fine-grained white sandstone, commonly less than 3 mm thick; the sandstone layers become iron-stained upon exposure. This type of bedding is not restricted to the Hungerford Slate but is also found in Skeels Corners and Morses Line slates. The slate in the Hungerford is poorly exposed, but limestone bioherms and local dolomites, which form only a small fraction of the total volume of the formation, crop out prominently. Bioherms are especially common in the lower quarter of the slate for a mile south of Cutler Pond, and about 35 feet stratigraphically above them is a prominent zone of arenaceous dolomite (Fig. 3, cols. 9–11).

THICKNESS: It is impossible to measure the thickness of the Hungerford in many of its outcrops, especially south of the Missisquoi River, because the upper limit is indeterminable. At the type locality (Fig. 3, col. 5) cleavage has obscured the bedding, but about 400 feet of beds seem to be present. The Hungerford is about 35 feet thick 2¼ miles south of Cutler Pond, but it increases to 400 feet at Cutler Pond and then diminishes again to an estimated 200 feet in a valley at the International Boundary, although no shales are exposed there (Fig. 3, cols. 7–11).

East of the St. Albans-Sheldon Springs road, slate assigned to the formation seems to range from 200 to 500 feet in thickness, but folding prevents exact measurement.

STRATIGRAPHIC RELATIONS: Rockledge Conglomerate grades upward in unbroken succession into Hungerford Slate. The contact can be seen at several places but is best exposed along the ridge 3½ miles north of St. Albans, just west of the road to Highgate Falls. Wherever the Rockledge is absent the Hungerford lies directly on Skeels Corners Slate, and the two formations are hard to differentiate.

The contact between Hungerford Slate and the Saxe Brook Dolomite is nowhere exposed, but the parallelism of structure within the two formations suggests that they are conformable.

In the Greens Corners slice (Fig. 2) the slates assigned to the Hungerford are believed to have been carried westward over the Morses Line Slate along the Oak Hill thrust.

FAUNA: Few fossils have been found in Hungerford Slate, and no lists or descriptions have been published although a study of the fauna is now in manuscript. The fauna includes the graptolite *Dictyonema schucherti* Ruedemann, 1933. The trilobites are *Catillicephala lata* (Raymond), 1937, and *Vermontella? minor* (Raymond), 1937. The brachiopods are *Lingulella franklinensis* and *Lingulepis vermontensis*. A problematicum is referred to *Pseudomatthevia?*. The fauna is assigned to the *Cedaria* zone.

Correlation of the Cedaria *Zone*

The discussion of the *Cedaria* fauna of the Marathon Uplift of Texas by Wilson (1954) appears to have resolved indirectly one of the most perplexing correlation problems in Vermont; namely, the proper placing of the boundary between the Middle and Upper Cambrian in Vermont. The St. Albans Shale has long been known as one of the few formations in the United States containing European paradoxidid trilobites. With such a fauna, its Middle Cambrian age has not been disputed, but the associated species are very much like those present in the overlying strata, (*see* faunal lists given above and Howell, 1937) which are assignable to the *Cedaria* zone. In the past the boundary between the Middle and Upper Cambrian has been drawn above the St. Albans Shale and below the Mill River Conglomerate (except on Cambrian Correlation chart (Howell *et al.*, 1944) where the line was drawn above the Skeels Corners Slate). This produces an unfortunate separation of faunas.

Wilson's Texas faunas have now shown a second association of *Cedaria* forms with European Middle Cambrian types, and he has come to the conclusion that the *Cedaria* zone may be correlative with the Middle Cambrian of Europe. Application of this correlation immediately solves the problem in northwestern Vermont by placing all the formations from St. Albans Shale through Hungerford Slate in the Middle Cambrian.

Upper Cambrian

Gorge Formation

NAME: The rocks included in Gorge Formation were first recognized as an independent unit by Keith (*in* Raymond, 1925, p. 137), who proposed to call them Missisquoi Formation, but as that name was already in use in central Vermont Schuchert (1937, p. 375) substituted the name Gorge Formation. Even after the first naming of the formation, rocks properly assignable to the Gorge were erroneously correlated with, or assigned to, Milton Dolomite and Mill River Conglomerate (Keith, 1932, p. 372).

DISTRIBUTION: The type locality of the formation is in the gorge of the Missisquoi River at Highgate Falls, where the river has cut deeply into the beds and made fine exposures. The type section (Fig. 3, col. 6) has been described by Schuchert (1937, p. 1052–1055). During the summer of 1954 Mary Gilman Clark re-examined the section in detail and made carefully spaced collections of fossils; because her revision of the section will be published later the problem will not be discussed at this time. Briefly, the type section consists of a basal dolomite, middle thin-bedded slate and limestones, and an upper slate which was assigned by Keith (Schuchert, 1937, p. 1076) to the Highgate Formation. In this report all rocks below the Highgate Falls thrust are assigned to the Gorge Formation.

North of the type section, only the lower dolomite is generally exposed. This unit forms a discontinuous ridge running northward from a point about 0.8 mile west-northwest of Highgate Center past the eastern side of Cutler Pond to the International Boundary (Pl. 1). East of this ridge lies a valley that may be underlain by unexposed middle Gorge Slate.

Limestones of the Gorge Formation are exposed at only two localities north of Highgate Center. The first exposes angular limestone breccias below bedded limestones in a small road-metal quarry 1.17 miles N. 35° E. of Bench Mark 308 in Highgate Center. The bedded limestone has yielded an undescribed fauna con-

taining *Idiomesus tantillus* and other forms suggesting a tentative correlation with Zones 2 and 3 of the type Gorge section (Raymond, 1924).

The second locality exposes a patch of limestone conglomerate on the east slope of the ridge of basal dolomite 1.55 miles N. 6° E. of Bench Mark 308 in Highgate Center.

THICKNESS: The lower dolomite of the Gorge is approximately 550 feet thick a mile N. 20° W. of Highgate Center, but it thins gradually northward to 425 feet at 45° N. Lat., then abruptly to 60–75 feet at the International Boundary (Fig. 3, cols. 6–11).

The overlying shale-limestone-shale sequence is present only at the type section (Fig. 3, col. 6), where it exceeds 400 feet in thickness. Northward it has been removed by erosion before deposition of the Highgate Formation.

The only outcrops south of the Missisquoi River that are correlated with the Gorge lie in the bed of Hungerford Brook about 0.6 mile above the mouth. These exposures are apparently part of the lower dolomite, but thick lake deposits in this area prevent tracing, so the correlation is uncertain. No rocks correlative with the Gorge have been recognized on the eastern side of the area.

STRATIGRAPHIC RELATIONS: Gorge Formation rests upon the deeply eroded upper surface of Hungerford Slate, and in the valley south of Cutler Pond at least 350 feet of Hungerford beds were removed before the Gorge was deposited; the channeled contact is especially well exposed in the cliff on the eastern side of Cutler Pond, where basal dolomites transect 30 feet of Hungerford Slate in a distance of 100 feet along the strike.

The Gorge is in normal contact with the Ordovician Highgate Formation only at the roadside quarry mentioned above, a mile N. 35° E. of Highgate Center and in a narrow strip, west of the Highgate Falls thrust, running 1.6 miles south from the International Boundary. Elsewhere the two formations are in fault contact. In the quarry, the Highgate lies on beds containing a Zone 2 and 3 Gorge fauna, indicating removal of much of the upper Gorge. At the International Boundary, north of the type section, only 20–200 feet of the lower Gorge dolomite remains (Fig. 3, cols. 10, 11), showing even more severe erosion than that seen northeast of Highgate Center.

FAUNA: Fossils of Gorge Formation have been described by Raymond (1924, 1925, 1937), Rasetti (1944, 1945, 1946), and Gilman (unpublished M.A. thesis, University of Wyoming, 1955), and they have been discussed incidentally by a number of other authors. Preliminary study has suggested that the *Ptychaspis-Prosaukia* zone of the Franconian and possibly the basal Trempealeauan is represented. This contradicts the dating made previously (Shaw, 1953, p. 140–141), wherein the *Hungaia magnifica* zone was stated to be latest Trempealeauan on the basis of what now seems to have been too much reliance on the evidence of the brachiopods. The fauna collected by C. H. Kindle (1936) above the Highgate Falls thrust and described by Shaw (1951, 1953) as coming from upper Gorge Formation and of latest Trempealeauan age is actually earliest Ordovician (*Bellefontia* zone) (Shaw, 1955a). In addition, a reconsideration of the structure has led the author to place the beds in question in lower Highgate Formation and not the Gorge.

CORRELATION: The middle Gorge limestones, containing the *Hungaia magnifica* fauna now seem to be correlative with the upper or upper middle Franconian, although precise zonal correlations must await further study. Within Vermont the Gorge limestones can be correlated with the Clarendon Springs Dolomite (Fig. 5), for each contains the *H. magnifica* fauna (Shaw, 1953, p. 140).

LOWER ORDOVICIAN

Introduction

In many ways the Ordovician rocks of the St. Albans area are the most difficult to study. The sequence is composed of two major units, a lower limestone (Highgate Formation) and an upper slate (Morses Line Slate), whose general outlines can easily be mapped but which are highly complex in detail. Close folding, well-developed axial plane cleavage, thrust faulting and extensive glacial cover all tend to complicate the study of both formations, and in Highgate Formation numerous unconformities tend to obscure the stratigraphic relations still further.

In previous publications the names Georgia Slate (Keith, 1923, p. 122 ff; 1932, p. 378–379; Schuchert, 1933, p. 379), Corliss Breccia (Schuchert, 1937, p. 1074), Grandge Slate (Schuchert, 1937, p. 1073–1075), and the erroneous name upper Gorge Formation (Shaw, 1951, 1953) have been used for the Ordovician rocks of the area. In this paper the name Georgia Slate is abandoned because Walcott's name for Lower and Upper Cambrian rocks should not be extended to the Lower Ordovician. The name Grandge Slate (pronounced Grahn-zhay) is dropped for reasons to be discussed below, and the upper Gorge beds are now included in the lowest Highgate Formation. Only Corliss Conglomerate is retained, as a member of Morses Line Slate.

Highgate Formation

NAME: This name is used here in a somewhat different sense from its original definition (Keith, 1923, p. 114–115). Keith's first description of the Highgate Slate included rocks now assigned to Highgate Formation as well as parts of Gorge Formation, Hungerford Slate and Skeels Corners Slate. The Highgate was dated as Upper Cambrian by Keith, but the Ordovician age was recognized by Raymond in 1924 (p. 462); this dating was not accepted by Schuchert until 1937 (p. 1072). Schuchert (1937, p. 1070 ff.) partly corrected Keith's errors of correlation by recognizing that Highgate Formation crops out only north of the Missisquoi River, but he erred in including certain Morses Line Slates in the Highgate. It was Schuchert who changed the name from Highgate Slate to Highgate Formation.

The Highgate, as here defined, also includes the beds formerly named Grandge Slate (Schuchert, 1937, p. 1073–1075). The type Grandge overlies a thin series of Highgate limestones with a marked angular unconformity (Schuchert, 1937, pls. 7, 8), but investigation of the Highgate elsewhere has revealed the presence of numerous angular unconformities throughout the unit, and inasmuch as it has proved impossible to map them or to use them as boundaries for separate formations, all of them, including the unconformity at Grandge Farm, are regarded simply as local features within Highgate Formation.

DISTRIBUTION: No type locality for the Highgate was chosen by Keith when he named the formation, but Schuchert (1937, p. 1070) chose "the upper sequence in the gorge of the Missisquoi River at Highgate Falls." Inasmuch as all the beds exposed below the Highgate Falls thrust are now referred to Gorge Formation Schuchert's definition is here interpreted rather freely, and the series of outcrops above the thrust and on the western edge of Highgate Center is regarded as the type locality. Thus interpreted, the formation lies in a strip half a mile to a mile wide trending about N. 20° E. from the type locality at Highgate Center. Near the International Boundary the outcrop is divided into a wide eastern band and a narrow western one by the Highgate Falls thrust (Pl. 1). East of the Highgate Falls thrust the formation is folded into a syncline, but except at Highgate Center the western limb of this Highgate syncline is faulted away. The thin succession under the Highgate Falls thrust dips eastward in regular continuation of the rest of the sequence on the western side of the area.

LITHOLOGY: The lowest Highgate beds lie above the Highgate Falls thrust in the north and east walls of Highgate gorge. These beds, inaccurately called upper Gorge formation by Shaw (1951, 1953), begin with a 10–15 foot bed of edgewise conglomerate containing slabby dolomitic limestone phenoclasts in an arenaceous dolomite matrix. This conglomerate has been attributed by Shrock (1948, fig. 168) to disruption of bedded limestone bottoms by unusually severe storms. This seems to be the best explanation because it accounts for the introduction of large amounts of sand into the matrix and explains the slabs that are free at one end although still connected at the other to undisturbed beds at the base of the conglomerate. Above the conglomerate are about 15 feet of black, crystalline, fossiliferous limestone in beds a quarter of an inch to 1 inch thick. These beds yielded the "upper Gorge fauna" of Shaw (1951, 1953). Equivalent beds are not present at the contact between Gorge dolomites and the Highgate at the International Boundary.

Except for this local exposure in the gorge the most commonly exposed rock in the Highgate is a banded blue limestone with beds ranging 1–6 inches thick, but in the valleys there are prob-

ably extensively developed calcareous slates. The upper part of the Highgate is predominantly argillaceous whereas the lower part is more calcareous, and it is likely that limestone and slate are present in the formation in roughly equal amounts, although the latter is less well exposed. Bioherms are rare; the largest, 1.96 miles N. 22° E. of Highgate Center, was described by Schuchert (1937, p. 1078). Intraformational limestone conglomerates are present at numerous stratigraphic levels, but they are of local extent.

UNCONFORMITIES: One of the most interesting features of the Highgate Formation is the presence within it of several angular unconformities. Most are exposed only locally, but one is traceable over a distance of at least 2 miles, where it is marked by lenses of a peculiar small-pebble conglomerate composed of rounded pellets of pale-blue limestone an eighth of an inch to half an inch in diameter and packed closely in a blue limestone matrix. The lenses are less than 10 feet thick and are discontinuous; the conglomerate may have been formed by the churning action of benthonic scavengers.

The southernmost exposure of this largest unconformity, which is here named the Rock River unconformity, is 2.00 miles N. 17° E. of Bench Mark 308 at Highgate Center. In this outcrop (Fig. 10) two other minor unconformities are present and are truncated by the Rock River unconformity. The Rock River unconformity is also exposed along the length of the knoll 0.44–0.80 mile N. 22° E. of the location shown in Figure 10 and again in the knolls 1.12 and 2.19 miles N. 22° E. of the first exposure. In all exposures the unconformity lies on the western side of the knolls and is accompanied by some development of the small-pebble conglomerate.

The unconformity at Grandge Farm has been described by Schuchert (1937, p. 1073–1074, Pls. 7, 8), 5½ miles N. 12° E. of Highgate Center. At this place bedded lower Highgate limestone is overlain by beds of conspicuously laminated gray calcareous slate, up to 4 inches thick, interbedded with gray limestone averaging an inch in thickness. Because the shale is more resistant than the limestone the bedding shows very well, but the weathered surface gives a false impression of the amount of clastics present. Schuchert did not realize that the Highgate Falls thrust is present at Grandge Farm and cuts off the beds just above the unconformity, so he estimated the "Grandge" Slate above the unconformity to be "at least hundreds of feet thick", but in fact it ranges from a thin veneer less than an inch thick to a maximum of 50 feet beneath the thrust.

Another angular unconformity in the upper part of the Highgate is exposed immediately east of the road south from Grandge Farm and 0.08 mile south of 45°00′ N. Lat.

Inasmuch as it is impossible to correlate these minor unconformities from one outcrop to another in all cases except Grandge Farm and the Rock River unconformity, they must for the present be included in Highgate Formation. The unconformities seem to be due in part to movements of the sea floor and in part to slumping although there is generally no brecciation or crumpling of the beds such as submarine slumping should cause (Schuchert, 1937, Pls. 7, 8).

THICKNESS: The extensive glacial cover, abrupt facies changes, and numerous unconformities within Highgate Formation make it impossible to measure detailed sections, but in general the section consists of a lower, primarily limestone unit 500–1000 feet thick, a middle shaly unit about 1500 feet thick, and an upper conglomerate and limestone unit that forms a thin capping in some outcrops along the Highgate Falls thrust near the International Boundary. These three units are gradational vertically through several hundred feet, and the thicknesses are only approximations.

AGE AND CORRELATION: The lower Highgate beds in Highgate gorge have yielded a fauna discovered by C. H. Kindle (1936) and described by Shaw (1951, 1953, 1955a). As has been pointed out on a previous page, this fauna was erroneously dated as Upper Cambrian (Shaw, 1951, p. 99–101) and placed in the upper Gorge Formation. The fauna is now recognized as Tremadocian and includes the following species: Trilobita: *Parabolinella triarthroides*, *Plicatolina kindlei*, *Highgatella gelasinata*, *Hardyoides glabrus*, *Symphysurina* (*Symphysuroides*) *minima*, *Missisquoia typicalis*, *Geragnostus* (*Micragnostus*) *bisectus*, *Pseudagnostus*

bilobus, *P. arancavelatus*, *Litagnostus raymondi*, and *Homagnostus* sp. Brachiopoda *Syntrophina carinifera*, *Archaeorthis crassa*, *Glyptotrophia spinosa*, *Apheoorthis* sp., and *Orthis* sp. Gastropoda: *Dirhachopea* sp. Cnidaria: *Serpulites acus*. Graptolithina: *Dictyonema vermontense*. Bryozoa: "*Heteronema*" sp.

The presence of *Symphysurina* (*Symphysuroides*) suggests correlation with either Zone A or B of the Garden City Formation of Utah (Ross, 1951, p. 29) and with the Gasconade of the typical Canadian. Physical relations also suggest that these beds are below the *Leiostegium*-bearing beds higher in the Highgate. The common absence of *Symphysurina* beds between the Gorge and the *Leiostegium* beds indicates that the *Symphysurina* beds are exceptionally limited.

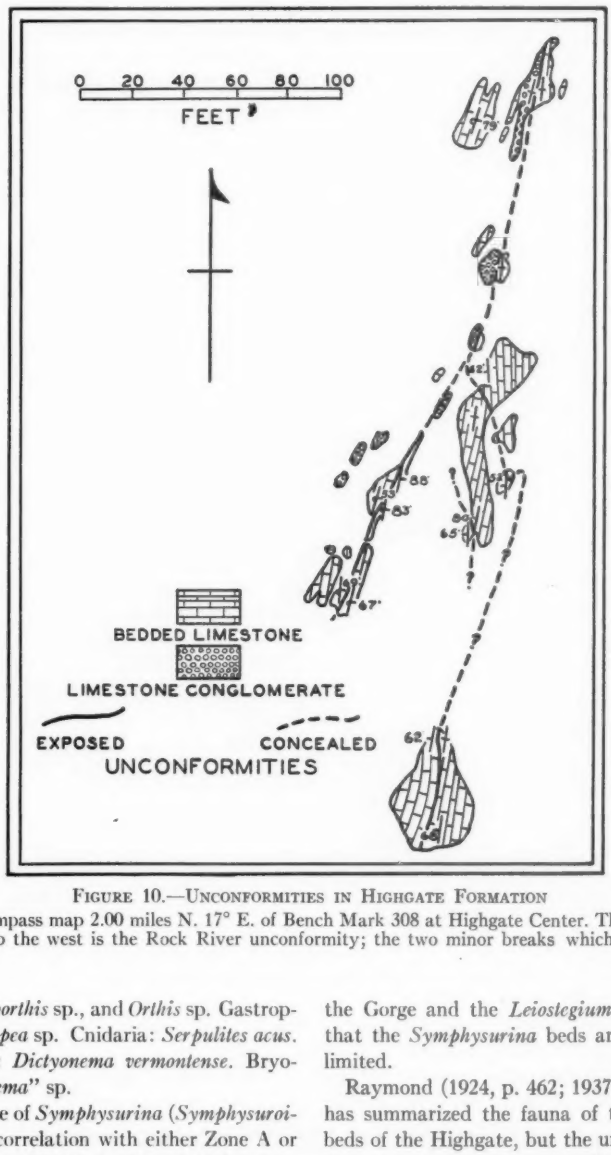

FIGURE 10.—UNCONFORMITIES IN HIGHGATE FORMATION

Tape-and-compass map 2.00 miles N. 17° E. of Bench Mark 308 at Highgate Center. The through-going unconformity to the west is the Rock River unconformity; the two minor breaks which it truncates are not named.

Raymond (1924, p. 462; 1937, p. 1133–1135) has summarized the fauna of the *Leiostegium* beds of the Highgate, but the unstudied collections from nearly 20 new localities show a large fauna and demonstrate that nearly every outcrop more than 2 miles north of Highgate Cen-

ter is fossiliferous. The heavy pygidia of *Leiostegium* are the most commonly identifiable fossils found in the Highgate, but in many places the limestones are a virtual coquina of macerated, unrecognizable trilobite and brachiopod fragments. Other than Raymond's summary, little is known of the large fauna of the *Leiostegium* beds of the Highgate, but the abundance of *Leiostegium* and the absence of *Symphysurina* warrant correlation with Zones D and E of the Great Basin Ordovician (Ross, 1951, p. 29; Hintze, 1952, p. 5, 9, 10).

Middle (?) Ordovician

Morses Line Slate

NAME: The name Morses Line Slate was proposed (Shaw, 1951, p. 98) for the wide expanse of slate and limestone conglomerate that underlies most of the central part of the St. Albans area. The name is taken from the village of Morses Line on the International Boundary (Pl. 1). The type locality may be regarded as lying southeast of Morses Line, where slates of the formation are well exposed. When discussed by earlier workers, these beds were referred either to Georgia Slate (Keith, 1923, p. 122; 1932, p. 378; Schuchert, 1933, p. 379) or to Highgate Formation (Schuchert, 1937, p. 1070–1071 and elsewhere).

The limestone conglomerate heretofore known as Corliss conglomerate, as restricted by Schuchert (1937, p. 1074), is also included in Morses Line Slate as a member. This conglomerate is widespread but differs only in length of outcrop from the many other conglomerates throughout the Morses Line. The name Corliss is not retained for the whole formation because it seems unwise to choose a limestone conglomerate as type for a slate unit.

DISTRIBUTION: Morses Line Slate underlies a large part of the center of the mapped area north of St. Albans Hill (Pl. 1). The western boundary of the formation runs almost straight N. 20° E. from that hill. The eastern side is hidden by rocks above the Oak Hill thrust. The exposure of the Morses Line widens northward and is about 5 miles wide at the International Boundary.

LITHOLOGY: Most of Morses Line is slate, both calcareous and non-calcareous. Thin-bedded limestone and limestone conglomerate form only local lenses, but these lenses are prominent because they resist erosion. Dolomite is present on the 320-foot hill half a mile south-southwest of Morses Line, but it was not seen elsewhere. The limestone conglomerates in the Morses Line are somewhat different from those in older formations. They are composed almost wholly of pale-blue biohermal fragments in a blue limestone matrix. The blocks range from small pebbles to a few masses nearly 100 feet long, but the mode is about a foot in diameter. There is little sand in either the matrix or the phenoclasts, and the lenses are associated in many places with bedded argillaceous limestone and calcareous slate. The lenses are rarely more than 50 feet thick and most crop out for only a few score yards, but the Corliss Conglomerate Member is exposed discontinuously for almost 2 miles north-northeast of the crossroads at elevation 383 a mile west of Greens Corners.

THICKNESS: It is impossible to make an accurate estimate of the thickness of the Morses Line because of glacial cover and small-scale folding and because the Morses Line is bounded, top and bottom, by thrusts. An estimated thickness of 2500 feet is probably a minimum. The formation thins southward, probably due in part to erosion and in part to faulting.

STRATIGRAPHIC RELATIONS: On the west Morses Line Slate is believed to be thrust over Highgate Formation and older rocks along the Gore thrust (Fig. 2), whereas on the east the formation is overridden by the Oak Hill thrust. Only at St. Albans Hill does the Morses Line appear to lie in sedimentary contact with Hungerford Slate (Pl. 1; Figs. 3, 15).

AGE AND CORRELATION: The first fossils found in the Morses Line were described by Raymond (1937, p. 1115–1118; 1140–1141) from pebbles in Corliss Conglomerate Member. The fauna includes *Symphysurina* (*Symphysuroides*) *latus* (Raymond), 1937, and *S.* (*Symphysurinella*) *corlissensis* (Raymond), 1937. (The other species listed by Raymond seem to be synonymous with these two). The presence of *S.* (*Symphysurinella*) indicates a Gasconade age for the pebbles and suggests a correlation with Zone C and the top of Zone B of Ross (1951, p. 29) and Hintze (1952, p. 5, 6–8) in the Great Basin. The only other fossiliferous outcrop thus far discov-

ered in the Morses Line is in a large gravel pit on the northwest flank of St. Albans Hill where a boulder in a limestone conglomerate that seems to lie at about the same stratigraphic level as Corliss member contains a molluscan fauna. Only one poorly preserved cephalopod could be collected, but it was sent to Dr. R. H. Flower, who has kindly furnished the following information about it: "I believe the specimen to represent an *Ellesmerocceras* and to be lower Canadian, but there is not enough there to eliminate the possibility of its being *Basslerocceras* and Middle Canadian in age." Thus, while a Lower Ordovician (Gasconade) age is shown for at least some of the pebbles in the conglomerates of the Morses Line, no information is now available to date the slate or the matrix of the conglomerates.

On August 26, 1954, Dr. W. M. Cady examined the Morse Line with the author and said that it closely resembles the Middle Ordovician (Clark, 1934, p. 6; Ulrich and Cooper, 1938, pl. 58) Stanbridge Slate of Canadian geologists. He advised retention of the name Morses Line until correlation with the Stanbridge could be proved. Use of the name Stanbridge at this time would suggest much greater certainty regarding correlation than actually exists. This correlation is markedly different from that given earlier (Shaw, 1951, p. 98) in which the Morses Line was stated to underlie Highgate Formation. The older correlation was based on the assumption that Corliss Conglomerate Member was intraformational and that the unfossiliferous matrix was, thus, of approximately the same age as the pebbles. If Dr. Cady's suggested correlation is correct, then the pebbles are older than the matrix, and the Morses Line is younger than the Highgate.

Problems for Investigation

The purpose of this study was primarily to determine the stratigraphic succession and to resolve the major structural complexities in the St. Albans area. Since the emphasis during the field work had to be placed on these two major problems many other interesting and important ramifications had to be neglected. Some of the most important of these secondary problems for which data are not now in hand include:

(1) Petrographic study of all sedimentary units to discover microscopic features that would aid in identification of their metamorphic equivalents to the east.

(2) Studies of facies changes, especially in the Dunham Dolomite, Skeels Corners Slate, Rockledge Conglomerate and Saxe Brook Dolomite.

(3) Intensive collection of fossils in the upper part of the lower Parker slate to determine whether the *Plagiura-Kochaspis* and *Wenckchemnia-Stephenaspis* faunas (Rasetti, 1951, p. 87–93) exist in Vermont and to locate the Lower Cambrian-Middle Cambrian boundary with more precision than is now possible.

(4) Thorough examination of the Highgate Formation to establish a stratigraphic and faunal succession within it so that the isolated segments of the formation can be correlated.

(5) Examination of the Morses Line Slate for fossils and tracing of the unit into Canada to attempt a correlation with the units recognized by Canadian geologists.

(6) Tracing and dating of equivalents of the Woods Corners Group north, east and south of the St. Albans area.

Structural Geology

General Statement

The St. Albans area, as defined in this report, lies near the northern end of the Rosenberg slice, one of a series of large thrust sheets lying east of the great Champlain overthrust. The Rosenberg slice was named from Rosenberg, Quebec, by Clark (1934), who briefly described the Canadian section. South of the St. Albans area the structure has been described by Cady (1945). In the St. Albans area the Rosenberg slice is folded into a broad, shallow northward-plunging syncline, whose eastern limb is broken by the Oak Hill, Fairfield Pond, and St. Albans thrusts (Fig. 2). Near the International Boundary several subsidiary faults break the syncline into minor slices. Four klippen lie on the eastern side of the thrust sheet (Fig. 2; Pl. 1).

Major Thrusts

Faulting is the most prominent structural feature of the area and is developed extensively

on both large and small scales. Five large overthrusts and countless smaller ones are recognizable. Normal and tear faults are of subordinate importance.

CHAMPLAIN THRUST: The Champlain thrust (Fig. 2) is the master fault in northwestern and west-central Vermont, and it has been described several times (Keith, 1923, p. 104; Cady, 1945, p. 565) since it was first recognized by Logan (1863, p. 281, 859). It can be traced from the vicinity of Middlebury, Vermont, northward for 50 miles, but it extends only a short distance into Canada before disappearing under other thrust sheets (Clark, 1934). Throughout most of its length the Champlain thrust is the westernmost fault of the western Vermont faulted belt, and it forms a natural boundary between the Ordovician rocks of the Champlain Valley to the west and the thrust-faulted and metamorphosed Cambro-Ordovician sequence to the east. North of St. Albans, however, the trace of the Champlain thrust swings eastward and to the west of it the Highgate Springs Series appears (Kay, 1958); north of Highgate Springs, the Philipsburg slice also appears west of the Champlain thrust.

The plane of the Champlain thrust is best exposed at Lone Rock Point west of Burlington in west-central Vermont (Schuchert, 1937, Pl. 3), where the dip is gently easterly, but it can also be seen in the St. Albans area at Fonda Quarry (the Smith's Lime Works mentioned by Logan), and the trace of the fault can be located to within 10 feet at many places. At Fonda Quarry the thrust is marked by a thick zone of gouge and does not show the well-defined plane seen at Lone Rock Point. The dip of the fault does not seem to exceed 15° E. and is probably flat at some places. It is impossible to make any reliable estimates of the net slip of the Champlain thrust, but if the stratigraphic throw is estimated at 10,000 feet and the dip of the fault averages 10°, the net slip, determined trigonometrically, would be slightly under 11 miles. The palinspastic map presented by Cady (1945, p. 568) indicates that he considered the net slip to be about 20 miles in the latitude of St. Albans. Like the other thrust faults in northwestern Vermont, the Champlain thrust appears to have rotated around a southern pivot so that the net slip increases northward.

SAXE THRUST: The Rosenberg slice is divided in its northern part by a number of small thrust faults, of which the most westerly and the smallest is the Saxe thrust, running through the western side of the Saxe Brook valley and thence probably northward to the International Boundary.

The evidence for this thrust is that the Middle Cambrian upper Parker Slate in Saxe valley lies only 200 feet above the Dunham Dolomite, whereas only 4 miles to the south correlative beds are about four times as far above the Dunham by addition of Lower Cambrian beds at the base of the formation. There seems little chance that this abrupt change is due to overlap because there is no evidence that the shore line was at the latitude of the Missisquoi River from Early until Middle Cambrian time. Thus, it appears that the basal part of the Parker has been faulted out. Secondary evidence for the Saxe thrust is a reversal of dip in the Parker on the west side of Saxe valley. This could be due to local folding, but coupled with other evidence, it seems to suggest drag-folding along a thrust.

The Saxe thrust is continued north to the International Boundary because the Parker does not reappear north of Saxe Brook valley, and the Saxe Brook Dolomite lies on the Dunham Dolomite, making a prominent westward facing scarp at the foot of which the trace of the fault has been drawn. The net slip of the Saxe thrust cannot be estimated directly because folding must have occurred before faulting. However, Skeels Corners Slate is about 2000 feet closer to Dunham Dolomite in Saxe valley than it is south of the Missisquoi River.

HIGHGATE FALLS THRUST: The second thrust east of the Champlain thrust has been recognized for many years, but it has never been described or mapped. It has been referred to briefly by Shaw (1951, p. 98; 1953, p. 138) as the Highgate Falls thrust, but no detailed description has been given. The fault is best exposed in the gorge of the Missisquoi River at Highgate Falls, but it extends beyond the International Boundary and also to the south (Pl. 1).

At Highgate Falls the base of the thrust sheet is a 10 to 15-foot bed of edgewise dolomite conglomerate at the base of the Highgate Forma-

tion. This bed forms the lip of the falls and is well exposed on both banks of the river. The fault plane is irregular, so that no representative dip measurements can be taken on it, but the average dips seem to be nearly flat in the gorge. Upstream from the bridge over Highgate Falls, however, the conglomerate dips 30° E. and the dip of the fault is assumed to steepen with it. Downstream, the fault plane steepens and rises to the west at 30°–40° and emerges from beneath glacial cover approximately at the road fork east of Carter Hill (Pl. 1). Gouge is not well developed along the fault plane and is not over 1½ feet thick, where observed. Although rocks directly beneath the thrust are badly contorted the effects of the movement are noticeable only for about 200 feet beneath the fault plane, and the bedded limestone at the mouth of the gorge is not deformed. The general structure of the footwall in the gorge is a syncline whose axial plane dips eastward at about 45° and probably strikes about N. 30° E.

The fault plane itself is only exposed in Highgate gorge, but the trace of the fault can be located to within 10–20 feet at several other places where it is marked by an abrupt change from the steep westerly dips of the overthrust block to the gently easterly dips of the footwall. For example, 2.2 miles N. 8° E. of Bench Mark 308 at Highgate Center the Gorge Formation below the fault strikes N. 21° E. and dips 34° E., and the slates of upper Highgate Formation above the fault strike N. 2° W. and dip 45° W. On the Grandge Farm, 5½ miles N. 12° E. of Highgate Center the lower Highgate (and the type Grandge Slate) dip 30° E. below the fault, and the upper Highgate dips 60°–70° W. above the fault. The fault plane is not exposed at Grandge Farm, but tracing of the edge of the overthrust block over the sloping ground suggests that the dip of the fault plane is about 20° E. Almost the entire thickness of Highgate Formation, probably about 3000 feet, is displaced along the fault.

It is clear from the fact that the Highgate Falls thrust places younger rocks over older that the Gorge and Highgate were folded before the faulting. The exposures in Missisquoi River gorge and at Highgate Center demonstrate this. Below the thrust, in the gorge, Gorge Formation is folded into a syncline, but above the thrust Highgate Formation is present in another syncline, whose axis diverges strongly from that below (Pl. 1). It would be impossible for an unfolded sequence to present the relations that now exist.

The southernmost outcrops in which the fault may be inferred lie northeast of the crossroads 4½ miles south of Highgate Falls on the road to St. Albans. Here, bedded limestone and limestone conglomerate believed to belong well up in Hungerford Slate are thrust over basal Hungerford. Here, too, the thrust shows the unusual relations caused by prior folding.

GORE THRUST: A short distance east of the Highgate Falls thrust there seems to be another fault of unknown magnitude which carries Morses Line Slate over Highgate Formation. The name derives from Gore Road, which runs north-northeastward from Highgate Center, and on which lie the North and South Gore Schools (Pl. 1).

West of Gore Road, Gorge and Highgate Formations dip steeply westward, vertically, or overturned steeply eastward on the east limb of the Highgate syncline (see below). East of the Gorge-Highgate exposures lies Morses Line Slate, strongly cleaved but having only moderate quaquaversal dips. The structural discordance between the two masses can be explained by faulting, which has brought the younger Morses Lines over the folded Highgate, or by assuming that this is a northern example of the great pre-Hortonville unconformity described by Cady (1945, p. 559–560) and later workers.

There is no direct evidence on which to base one's choice, but the first is selected because it can be explained by the same sequence of events already noted as accounting for the Highgate Falls thrust. The second hypothesis requires (1) folding of the Gorge-Highgate, (2) erosion of the section, (3) overlap of the Morses Line sea, and (4) a second period of thrusting (since Morses Line is itself overridden by older rocks to the east). The arguments need not be labored further; there is so little direct evidence that either choice is mainly a matter of opinion.

South of the Missisquoi River the course of the Gore fault and its relations to the Highgate Falls thrust are largely conjectural, and the in-

terpretation shown on Plate 1 is but one of several that could be true.

HIGHGATE SLICE: The fault slice of westerly dipping Highgate beds between the Highgate Falls and Gore thrusts is here called the Highgate slice. At Highgate Center the structure is broadly synclinal (see Highgate syncline below), but a mile north of Highgate Center the beds dip generally westward or are overturned toward the east across the entire slice. The overturning is proved by cleavage-bedding relations.

OAK HILL THRUST: This poorly exposed thrust fault is indicated chiefly by the structural pattern along the eastern side of the mapped area. The Oak Hill thrust was first recognized in Canada by Clark (1934, p. 4, Fig. 2; 1936, p. 136, Fig. 1, p. 141, Fig. 3), who indicated it in several figures but did not formally describe it. He traced the fault into the northeastern corner of the St. Albans area whence it has been traced southward during the present survey to the vicinity of St. Albans, where it seems to disappear.

The fault trace is recognizable principally by: (1) the abrupt change in the degree of metamorphism across the fault, (2) the development of cleavage in the foot wall, (3) by local reversal of dips, and (4) omission of beds. These features are well displayed along the St. Albans-Sheldon Springs road between 44° 50′ N. Lat. and 44° 54′ N. Lat.

East of the fault trace well-bedded slates and thin argillaceous limestones assigned, because of their stratigraphic position, to the Hungerford Slate are thrown into small-scale folds and are cleaved but not to the same extent as in the mashed and crumpled Morses Line Slates below the fault, in which commonly all trace of bedding has been destroyed. Between East Highgate and Bridgeman Hill, the actual trace Oak Hill thrust is covered by glacial debris, but it can be located in a general way because it lies between the gently folded structures of the Cambrian clastic sequence to the east and highly cleaved Morses Line Slate to the west.

Near Bridgeman Hill it is very difficult to trace the fault because additional intense cleavage had developed beneath the Bridgeman klippe and obscured the changes visible elsewhere across the fault. However, slight changes in the lithology of the slates suggest that the trace of the Oak Hill thrust runs just west of the klippe. North of Bridgeman Hill the trace of the fault can be located only in the most general way, for it appears that the overriding slates are as much cleaved as those overridden. The fault crosses the International Boundary about 0.6 mile west of 72° 55′ W. Long.

The amount of displacement and the dip of the Oak Hill thrust are not directly observable, and because of the poor exposures no significant estimates can be made.

The recognition that the southern end of the Oak Hill thrust lies in the St. Albans quadrangle renders the use of the name to the south by Cady (1945, p. 566–567) incorrect, but it is not within the province of this discussion to suggest what name should be applied to Cady's "Oak Hill thrust", although it may be that it correlates with the St. Albans thrust of this paper.

In criticizing the manuscript of this paper, Dr. Cady emphasized that the map pattern explained on Plate 1 by means of the Gore and Oak Hill thrusts might be explained by sedimentary overlap. The problem is by no means solved, and this alternative should be considered by future workers. Evidence supporting the belief that the contacts shown here are sedimentary and not faulted is as follows:

(1) West of St. Albans Hill, Morses Line appears to be directly on Hungerford Slate.

(2) From Bridgeman Hill south along the eastern boundary of the area Hungerford Slate is gradually cut out, as if by an erosion surface.

(3) Structures within Morses Line Slate are not related to those in the older rocks.

(4) If Morses Line is a Stansbridge equivalent it would be young enough to overlie the pre-Hortonville unconformity (Cady, 1945, p. 559–560).

The facts and inferences which tend to favor the thrusting hypothesized in this paper are:

(1) West of the contact between Morses Line Slate and the older rocks on the western side of the area the Morses Line is uniformly more highly cleaved than the older beds, except where both sequences have suffered extreme pressure, as near Bridgeman Hill klippe.

It is easy to account for this difference if the Morses Line forms the foot wall of a thrust, but it is difficult to account for the fact that beds above an unconformity are more metamor-

phosed than the older, underlying strata. This is especially true since both sequences are dominantly black slate and should, thus, react to uniform diastrophic forces in a comparable manner.

The Fairfield Pond thrust forms the eastern boundary of the mapped area northward of a point a mile south-southeast of Greens Corners. The fault plane is nowhere exposed, but it is assumed to lie at the base of the 400-foot ridge

FIGURE 11.—WEST-TO-EAST STRUCTURE SECTION HALF A MILE SOUTHEAST OF GREENS CORNERS
Csc = Skeels Corners Slate. Cp = Parker Slate. Cd = Dunham Dolomite. Cgi = Gilman Quartzite. Stippled area below Fairfield Pond thrust is Gilman Quartzite.

(2) Cleavages on either side of the eastern contact are unlike in attitude. The dip of cleavage in Morses Line Slate ranges from 36° to 85° E., and the mode is 70°; the mean is 65°. The slates east of the contact range from 25° to 59° E., with a 45° mode and a 40° mean. More than two-thirds of the dips in Morses Line Slate exceed the maximum dip noted east of the contact. The fact that both sets of cleavage are related to minor folding strengthens the idea of structural discontinuity.

(3) Wherever the dips can be seen along the eastern contact Morses Line Slate dips gently eastward (25° or less). There is no evidence of folding up of the beds, which would be necessary if these beds at one time passed up and over the topographically higher Cambrian rocks to the east. If there was no faulting there must have been folding, and none is in evidence.

GORE SLICE: That segment of the Rosenberg slice lying between the Gore and Oak Hill thrusts may be called the Gore slice. It contains the main outcrop of the Morses Line Slate.

FAIRFIELD POND THRUST: The Fairfield Pond thrust (Fig. 2; Pl. 1) was named and described by Booth (1950, p. 1163–1164), who recognized the evidence for its presence east of Greens Corners, but who may have described two separate faults under a single name. The southern part of Booth's original Fairfield Pond thrust is described below as the St. Albans thrust.

of Gilman Quartzite that forms Minister and Rice Hills and the unnamed ridge northeast of Greens Corners—the series of ridges that Booth has called the Oak Hill Escarpment. The best evidence for the existence of the fault may be observed half a mile southeast of Greens Corners (Fig. 11), where the vertical beds below the Fairfield Pond thrust contain approximately 600 feet of gray Dunham Dolomite overlying Gilman Quartzite, while to the east on the slope of the hill there are at least 1300 feet of eastward-dipping, notably limy Dunham. It does not seem possible to explain the changes in thickness or in lithology except by the faulting of the Dunham over itself at this point. This interpretation differs from that given by Booth (1950, p. 1163).

The northward extension of the fault as shown on Plate 1 cannot be proved by direct observation, but the following evidence seems to warrant it as at least a probable inference: (1) North of Greens Corners there is a deep valley formed in Gilman Quartzite and Dunham Dolomite, the two most resistant formations in the section. It seems unlikely that a continuous valley would form along the entire length of outcrop of these formations unless a zone of weakness, such as a fault would produce, determined its path. (2) The eastern wall of this valley is in thrust relation at Greens Corners (Fig. 11), and there is nothing observable northward to suggest that this relation

changes. (3) The rocks of the eastern wall of the valley remain constantly in a highly cleaved chloritic facies of Gilman Quartzite, but the west wall of the valley is formed by beds at various levels in both Gilman and Dunham. This lithologic discordance between the two sides of the valley is also paralleled by structural discordances as, for example, west of Sheldon Springs, where the folds in the Gilman west of the valley do not seem to be present in the Gilman east of the valley.

Save for the course of the fault east of Browns Corners and near Sheldon Springs the trace of the Fairfield Pond thrust shown on Plate 1 agrees closely with the map of Booth (1950, Pl. 1); there seems to be no evidence of the fault running over the ridge 2 miles west of Gates Hill, and on Plate 1 it is continued in the valley east of the ridge.

The net slip of the Fairfield Pond thrust cannot be measured directly. There is no evidence of this fault in the southern part of the St. Albans quadrangle.

ST. ALBANS THRUST: The trace of the St. Albans thrust forms the eastern boundary of the mapped area from its southern edge northward to a point a mile S. 10° E. of Greens Corners, where the thrust swings east and is replaced along its strike by the Fairfield Pond thrust. Because the trace of the St. Albans thrust runs eastward into the area mapped by Booth (1950) it was not mapped during this survey.

The clearest evidence for the St. Albans thrust may be seen a mile south and southeast of Greens Corners where the Dunham Dolomite above the Fairfield Pond thrust disappears beneath a 10- to 20-foot scarp of Gilman Quartzite lying above the Dunham in thrust relation. This relationship was shown on Booth's map, but he did not separate the two faults. It may be noted that Booth (1950, Pl. 1, and p. 1163) has mapped a series of small thrust slices north of the road east from Greens Corners. This area was studied and seems to be simply a badly cleaved succession of alternating quartzites and argillaceous quartzites; the cleavage seems to have been induced by the movement of the St. Albans thrust.

The fault plane of the St. Albans thrust is nowhere directly exposed, but the Gilman southeast of Greens Corners drops down over the Dunham in a manner that suggests a gentle westerly dip, probably induced by later folding. Such late folding is also indicated in the klippen. If the three southern klippen are remnants of the St. Albans thrust sheet, a minimum net slip of 1 – 1½ miles may be postulated along the fault; no maximum limit can now be set. It is possible that the two faults recognized here as the Fairfield Pond and St. Albans thrusts are, in fact, only one fault. However, since there are two faults present at Greens Corners and since the eastern one can be seen in the outcrop to cut across the western, it has seemed best to consider them separately until the northern end of the St. Albans fault is traced back into the Fairfield Pond thrust.

GREENS CORNERS SLICE: The name Greens Corners slice is here proposed for the newly recognized structural unit lying between the Oak Hill and Fairfield Pond thrust faults north of Greens Corners. South of Greens Corners the slice is bounded by the Oak Hill and St. Albans thrusts, but near the southern edge of the St. Albans quadrangle the Greens Corners slice merges with the Rosenberg slice because of the disappearance of the Oak Hill thrust (Pl. 1; Fig. 2).

Structurally, the Greens Corners slice is the isolated eastern, westward-dipping limb of the St. Albans synclinorium. Booth (1950, p. 1163) described the rocks here included in the slice as a "rather extensive though secondary anticline", but the over-all dips are to the west or overturned toward the east. The Gilman Quartzite commonly lies on the eastern side of the slice, but south of Greens Corners the Gilman, Dunham, and Parker successively disappear beneath the St. Albans thrust. Skeels Corners Slate has yielded *Bovicornellum vermontense* east of St. Albans Hill, and beds resembling the Hungerford Slate overlie the Skeels Corners.

Klippen

Five klippen are mapped on Plate 1; 4 lie on the eastern side of the area and are here named, from north to south, the *Bridgeman klippe*, the *Aldis klippe*, the *St. Albans klippe* and the *Johnnycake klippe*. The fifth is west of the Champlain thrust and is called the *Swanton Junction klippe*.

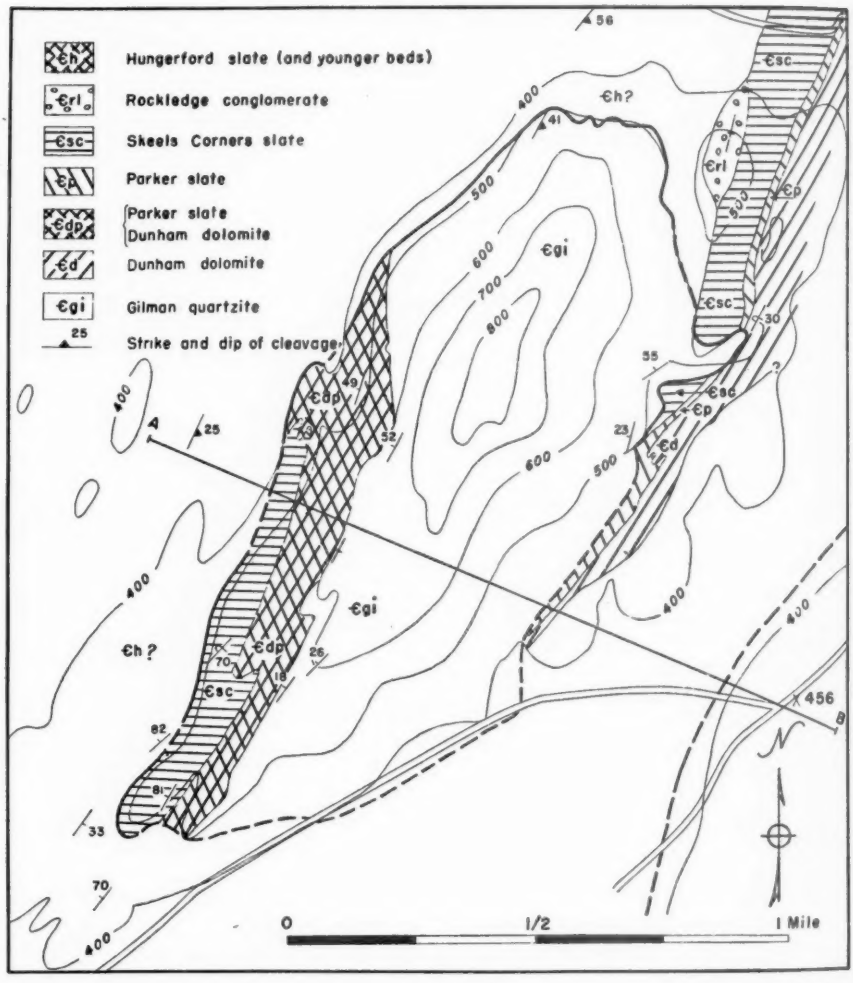

FIGURE 12.—LARGE-SCALE MAP OF THE BRIDGEMAN HILL KLIPPE

Base from U. S. Geological Survey topographic map of the Enosburg Falls quadrangle. Trace of the Oak Hill thrust west of the klippe has not been shown (*see* Pl. 1). Line of section A-B shown in Figure 13.

SWANTON JUNCTION KLIPPE: A quarter of a mile north-northwest of Swanton Junction erosion has isolated a small block of Dunham Dolomite from the main block of the Rosenberg slice to the east. Klippen are uncommon along the northern Champlain thrust so this one is named in spite of its small size.

BRIDGEMAN KLIPPE: Bridgeman Hill, in the northeastern corner of the mapped area, is formed by the largest klippe in the area, the Bridgeman (Figs. 12, 13), which is essentially a flat plate of Lower and Middle Cambrian rocks lying on the upturned edges of beds of the same age in the Greens Corners slice. The klippe has an area of about 1⅛ square miles, of which resistant Gilman Quartzite underlies about 80 per cent. Above the Gilman, on the west side of the klippe are dolomite beds of typical Dun-

ham lithology, and at the southern end of the hill are a few patches of slate and arenaceous dolomite that may be Parker Slate. These Parker (?) beds are so limited areally, however, that they are included in the Dunham in Figures 12 and 13 and on Plate 1. Above these slates lie well-bedded dolomitic shales that are referred to Skeels Corners Slate on the basis of lithology and stratigraphic position. No fossils or traces of them have been found in any formation in the klippe. Bedding on the eastern side of the klippe dips gently to moderately westward, but the dip steepens on the western edge, where vertical and overturned beds are found. The plane of the fault is exceptionally well exposed for more than a mile around the northern end of the klippe, where the highly cleaved slates below the fault have been eroded away, leaving good exposures. The plane is somewhat irregular, and shows a vertical "relief" of as much as 25 feet in 50 feet along the trace, but its over-all attitude is a gentle westerly dip of about ¾°. On the small spur on the eastern side of Bridgeman Hill the fault plane is exposed standing vertically, showing that there has been folding since the thrusting occurred. Cleavage is prominent both above and below the fault plane, but at no place was a gouge zone more than a foot thick observed.

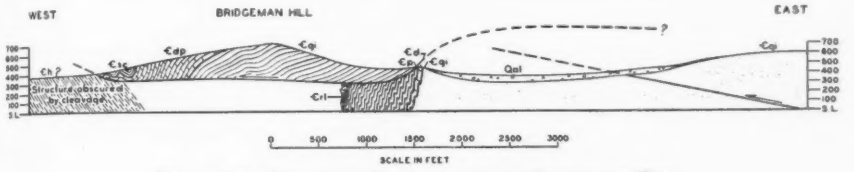

FIGURE 13.—STRUCTURE SECTION THROUGH BRIDGEMAN HILL
Formation symbols as in Figure 12. Location shown on Figure 12.

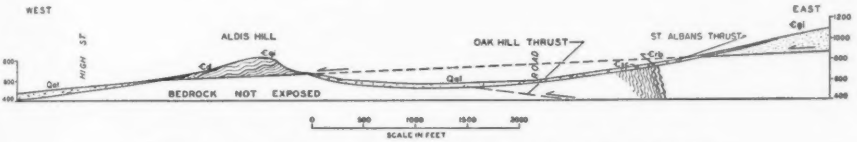

FIGURE 14.—STRUCTURE SECTION THROUGH THE ALDIS HILL KLIPPE
Qal = alluvium, ground moraine, and beach deposits. Ꞓsc = Skeels Corners Slate. Ꞓrb = Rugg Brook Dolomite. Ꞓd = Dunham Dolomite. Ꞓgi = Gilman Quartzite.

ALDIS KLIPPE: The Aldis klippe (Fig. 14) forms Aldis Hill in the northeastern part of the City of St. Albans. It is essentially a smaller version of the Bridgeman klippe, and most of it is made of Gilman Quartzite, overlain along the western side by small patches of Dunham Dolomite. Keith (*in* Longwell, and others, 1933, Fig. 14, p. 63; *in* Schuchert, 1937, Fig. 2c, p. 1006) presented a structure section through Aldis Hill, and he recognized its overthrust nature, but he believed that both the bedding in the klippe and the fault beneath it dipped eastward, whereas both actually dip westward. The only eastward-dipping beds on Aldis Hill are exposed along the highway that crosses its northern end; elsewhere westerly dips are clearly visible.

ST. ALBANS KLIPPE: This name is given to a small patch of cleaved, massive, chloritic Gilman Quartzite exposed in the southern part of St. Albans (Pl. 1). The outcrop is surrounded by glacial debris so that the nature of the underlying rocks cannot be determined directly, but slates provisionally assigned to Morses Line Slate are exposed 2000 feet north in the bed of Stevens Brook. The latter exposure also proves that the Aldis and St. Albans klippen are not connected.

JOHNNYCAKE KLIPPE: The Johnnycake[1]

[1] The name *Johnnycake* is taken from the correct name of the hill labeled "St. Albans Hill" on the St. Albans quadrangle topographic map. The name St. Albans Hill is applied locally to the first saddle north of Bellevue Hill, through which passes the road from St. Albans to Swamp School. For convenience, the name on the topographic map has

klippe (Fig. 15) is a synclinally folded thrust mass that lies on the eastern slopes of St. Albans Hill, south of St. Albans. As in the other klippen, the main mass of the thrust block is Gilman Quartzite, but on the western edge overturning brings down Dunham-like

eastward in order to reach its assumed roots on the west flank of Bellevue Hill (Fig. 15).

The foot wall of the thrust is better shown under the Johnnycake klippe than under any of the others. It is a tightly folded anticline whose axial plane dips eastward (Fig. 15).

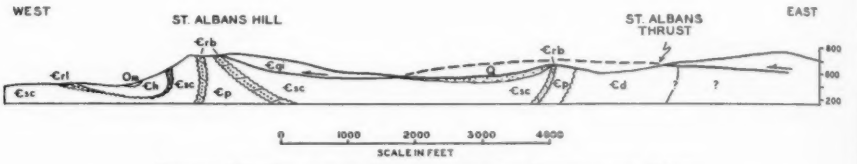

FIGURE 15.—STRUCTURE SECTION THROUGH JOHNNYCAKE KLIPPE

N. 65° W. through the crest of St. Albans Hill. Cgi = Gilman Quartzite. Cd = Dunham Dolomite. Cp = Parker Slate. Crb = Rugg Brook Dolomite. Csc = Skeels Corners Slate. Crl = Rockledge Conglomerate. Ch = Hungerford Slate. Om = Morses Line Slate. Wavy contacts are erosion surfaces. No vertical exaggeration.

dolomite, although in such small amount that it cannot be shown on Plate 1. The western edge of the klippe is also intricately faulted on a small scale so that Dunham and Gilman types of rock are cut out and repeated in confusing fashion. Lack of a sufficiently accurate base map made tracing of these faults impossible.

The trace of the St. Albans thrust can be followed easily across the top of St. Albans Hill. It is exposed on the eastern side of a northward-trending saddle at an altitude of about 820 feet where the Gilman overrides Parker Slate. Elsewhere to the north and south Rugg Brook Dolomite is overridden by the fault. The fault is marked by a gouge zone at least 3 feet thick and containing numerous fragments of dolomite; the gouge obscures the dip of the fault plane, but the overhanging Gilman suggests that it must be easterly and shallow. At the northern end of the hill a bed of massive chloritic Gilman Quartzite at least 50 feet thick has been undercut at several places, forming caves 1–6 feet deep; these caves seem to be weathered in the fault, whose dip here is slight. The eastern side of the klippe is buried under glacial deposits for most of its length.

Like the Bridgeman klippe, the Johnnycake klippe seems to have been folded subsequent to the thrusting, for the fault plane must rise

Exposed on the crest of the hill are bedded calcareous sandstones of Rugg Brook Dolomite. South of the crest erosion has breached the Rugg Brook and exposed the Parker in a trough-like valley that opens southward. The Parker is cut off to the south by a minor fault that places it against Skeels Corners, the formation that forms the steep west face of St. Albans Hill. On the western and northern foot of the hill discontinuous lenses of limestone conglomerate are believed to be Rockledge, but they are so badly cleaved and sheared that they no longer look like typical Rockledge, and they yield no fossils.

Near the foot of the west-central face of the hill there is a patch of Ordovician limestone conglomerate (*Om* in Fig. 15). The age is demonstrated by the fossils in the pebbles; it is also clear that the outcrop lies too near the Rockledge, both east and west, for there to be a full section present in either direction. The interpretation favored (Fig. 15) explains the situation by deep erosion of the Hungerford Slate and deposition of Morses Line Slate on it.

SOURCE OF THE EASTERN KLIPPEN: It seems logical to assume that the Aldis, St. Albans, and Johnnycake klippen are remnants of the St. Albans thrust sheet because they are so close to one another and are essentially alike lithologically. Likewise, it seems reasonable to suppose that only the St. Albans thrust could have been responsible for their emplacement because the Oak Hill thrust ends east of St.

been used throughout this report, although it is incorrect.

Albans Hill, and the Fairfield Pond thrust is covered by the St. Albans thrust sheet in this area. The Bridgeman klipp might have been emplaced along the Fairfield Pond thrust, but the structural and lithic similarities among the klippen argue strongly for a common origin.

Smaller Faults

SAXE BROOK TEAR FAULT: A strike-slip tear fault has developed in the Saxe Brook Dolomite at the north end of the Saxe valley (Pl. 1). It is reflected in areal plan by a widening of the outcrop and by dragging of the beds near the break. It is possible that the disruption of the formation is due to a sharp fold rather than a tear fault, but the visible deformation of the bedding seems too little to account for the amount of displacement by folding alone.

MELVILLE LANDING SLICE ZONE: Three-quarters of a mile south of Melville Landing, in the southwestern corner of the mapped area, Dunham Dolomite is fractured into several thrust slices which, although of small displacement, can be shown on the areal map (Pl. 1). These slices appear to have formed at a short distance above the Champlain thrust in response to the shearing stresses set up by movement of the main fault.

MINOR THRUST FAULTING: The minor thrust faults are literally countless in northwestern Vermont, particularly in shale formations. Small faults with offsets of a fraction of an inch to several feet are visible in nearly every slate outcrop. It is impossible to plot these minute breaks on the accompanying map, but the total displacement along them must aggregate hundreds of feet. This introduces a complicating factor into the problem of estimating thickness of stratigraphic units for which no compensation can now be made. These minor thrusts dip eastward at angles ranging from nearly flat to about 60°, but the vast majority dip less than 25°; the fault planes strike east of north in most cases, and the upper block seems invariably to have moved westward over the lower.

NORMAL FAULTING: Normal faulting is not well developed in the area, but a graben has been mapped in Dunham Dolomite half a mile south of Swanton Junction (Pl. 1). Another northwestward trending graben was noted in the Dunham a mile S. 25° W. of Tuller School, northwest of St. Albans, but it was too small to appear on the areal map.

POST-PLEISTOCENE FRACTURING: Fracturing of glaciated surfaces was noted on the west side of a small saddle 1.98 miles N. 36° E. of Browns Corners in the northwestern part of the Enosburg Falls quadrangle, where iron-stained, gray Morses Line Slate has been displaced along flow cleavage which strikes N. 40° E. and dips 55° S.E. One scarplet, which is traceable for 30 feet, has the eastern block raised about a quarter of an inch above the western. Along another fracture the eastern block stands half an inch above the western, but restoration of the weathered surface of the elevated side indicates an original displacement of about three-quarters of an inch.

St. Albans Synclinorium

The Rosenberg slice in the St. Albans area is folded into a northward-plunging asymmetric syncline 3–7 miles wide, with a gently dipping west limb and a vertical or overturned east limb. The syncline has been broken by the several faults described above. In accordance with the practice instituted by Cady (1945), who called similar large folds "synclinoria", the fold north of St. Albans Hill is here named the St. Albans synclinorium.

Within the St. Albans synclinorium folding is present on two intergrading scales. First are the large folds that are extensive enough to be mapped (Pl. 1; Figs. 8, 16). Several of these have been given names and are described individually below. Second, are the minor folds that have developed extensively in the argillaceous formations and less commonly in the bedded limestones. Heavy dolomites, like the Dunham, show little sign of minor folding even on the eastern limb of the synclinorium, where the beds are vertical or overturned. Thus, the scale of folding is dependent more on the competence of the beds than on the absolute amount of compression or upon the position of the bed within the synclinorium.

SKEELS CORNERS AND HUNGERFORD BROOK ANTICLINES: Two anticlinal folds in the west limb of the synclinorium involving the Skeels

Corners, Rockledge and Hungerford have been mapped (Pl. 1) half a mile south of Skeels Corners and north of Skeels Corners to Hungerford Brook; the southern fold is called the Skeels Corners anticline and the northern, the Hungerford Brook anticline (Fig. 9). Both are in the form of large "drag-folds" that formed parallel to the main fold of the synclinorium.

HIGHGATE SYNCLINE: The Highgate syncline is an open, shallow fold in the Highgate slice. It begins at the type locality of the Highgate formation on the western outskirts of Highgate Center. The axial plane strikes about N. 5° W. and dips about 45° E.; the plunge of the fold is gently north. The western limb dips 10°–25° E. and is exposed for a mile west and northwest of Highgate Center before terminating against the Highgate Falls thrust. The eastern limb dips steeply westward and is locally overturned. It can be traced with moderate certainty to the International Boundary.

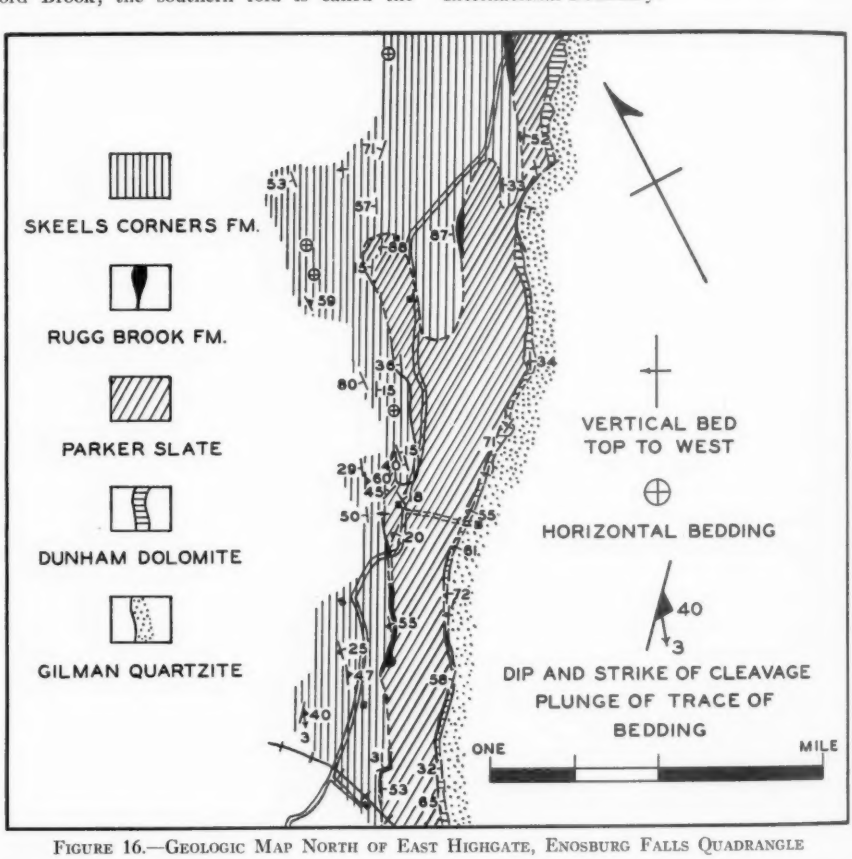

FIGURE 16.—GEOLOGIC MAP NORTH OF EAST HIGHGATE, ENOSBURG FALLS QUADRANGLE

MINOR FOLDS NORTH OF EAST HIGHGATE: Figure 16 is a map of the region north of East Highgate in the Enosburg Falls quadrangle. It is presented as one of the best examples of the intricately folded structures on the eastern side of the St. Albans synclinorium. In this area the key beds are Rugg Brook Dolomite, here developed as a bedded sandstone with carbonate cement, and Dunham Dolomite which, although thin, clearly marks the base of Parker Slate; Skeels Corners Slate is intricately folded throughout the area.

DRAG-FOLDING: Drag-folding is extensive in

argillaceous formations of the area. It parallels both the large- and small-scale structures faithfully and presumably developed at the time of the major folding. It is very useful in determining local structures, for the attitude of the drag-folds can be relied upon as a guide to

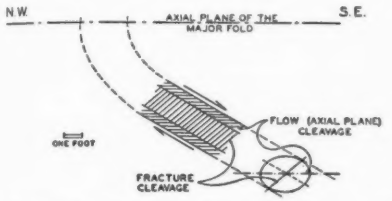

FIGURE 17.—FLOW AND FRACTURE CLEAVAGE IN DUNHAM DOLOMITE

Relations to bedding 1 mile S. 70°W. of Sheldon Springs.

the attitude of beds, plunge of major structures, and the direction to the adjoining major structure.

CLEAVAGE

DISTRIBUTION: Cleavage is found in rocks of all types in the area, but it is better developed in the eastern part of the St. Albans area than in the western and in the argillaceous rather than in the calcareous rocks. Flow cleavage is overwhelmingly dominant, although one instance of well-developed fracture cleavage was noted and is described below. Locally, intense flow cleavage has destroyed bedding in some of the slates, and in the eastern half of the area it has destroyed virtually all the fossils.

FLOW CLEAVAGE: The flow cleavage dips eastward at various angles from nearly horizontal to vertical; the mode is 50°–60° E. The strike is almost invariably east of north and veers from N. 15° E. to N. 30° E. in the southern part of the area to N. 40° E.–N. 60° E. in the northeast. The strike of the cleavage follows the strike of the formations in general, but the two depart from each other in detail.

The planes of the flow cleavage lie parallel to the axial planes of the minor folds throughout the area. This is an invaluable aid in determining the structure, for the relationship between the cleavage and bedding may be used to determine both the direction to the nearby folds, the direction and amount of plunge of fold axes, and the tops of the beds.

Fracture cleavage has developed in overturned Dunham Dolomite a mile S. 70° W. of Sheldon Springs, in the Enosburg Falls quadrangle where the bedding in the Dunham dips 30° S.E. and the fracture cleavage 40° N.E. The flow cleavage was developed only in a few thin shaly partings and it could not be used with assurance to determine whether the beds were in normal or inverted position. Figure 17 shows the relationships; the relationships between the better-developed fracture cleavage and the bedding showed that the beds were inverted. Later mapping substantiated this conclusion.

Refracted cleavage has developed in both the Hungerford and Morses Line slates, but it is uncommon. The origin of this feature is not certain; it may have been caused either by the development of flow cleavage in the weaker beds and fracture cleavage in the stronger or by the displacement of a single flow cleavage system by slight later movements of the resistant beds; inasmuch as slight movements postdating the major deformation are indicated by other lines of evidence, such as the warping of Bridgeman and Aldis klippen, the second explanation seems the more likely in the St. Albans area.

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INFOLDED METASEDIMENTARY ROCKS NEAR THE AXIAL ZONE OF THE CATOCTIN MOUNTAIN-BLUE RIDGE ANTICLINORIUM IN VIRGINIA

BY EDWIN O. GOOCH

ABSTRACT

A belt of metasedimentary rocks about 60 miles long and from a half to 2 miles wide occurs in the northwestern Piedmont of Virginia. It lies along or near the axis of the Catoctin Mountain-Blue Ridge anticlinorium about midway between the Late Precambrian-Early Cambrian formations in the Blue Ridge and Late Precambrian and possibly younger formations in the Piedmont.

The belt of rocks occurs as an infold that is locally faulted. The infold is overturned toward the northwest in the southwestern end but becomes open to the northeast. The metasedimentary rocks are equivalent in part to the Lynchburg and Swift Run formations.

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INTRODUCTION

An infolded belt of metasedimentary rocks about 60 miles long and from a half to 2 miles wide extends from near Batesville northeastward to near Ben Venue in Virginia (Fig. 1). It lies about 2 miles southeast of the Blue Ridge and is surrounded by the older Precambrian rocks of the basement complex. It occurs along or near the axis of the Catoctin Mountain-Blue Ridge anticlinorium about midway between the Late Precambrian-Early Cambrian formations in the Blue Ridge and Late Precambrian and possibly younger formations in the Piedmont. The size, location, and relationships of the belt indicate that it is an important element in the regional geology. Part of this remarkable belt has been noted by previous geologists (Geologic Map of Virginia, 1928; Bloomer 1950, p. 753–783), but the belt as a whole was first mapped and described by the writer.

ACKNOWLEDGMENTS

The work was done under auspices of the Virginia Division of Geology, and the writer is indebted to Mr. William M. McGill, State Geologist. A. A. Pegau, R. O. Bloomer, and H. J. Werner spent considerable time in the field with the writer and offered many valuable suggestions. The paper has been read by R. O. Bloomer, W. R. Brown, P. B. King, A. A. Pegau, and H. J. Werner. To each of these

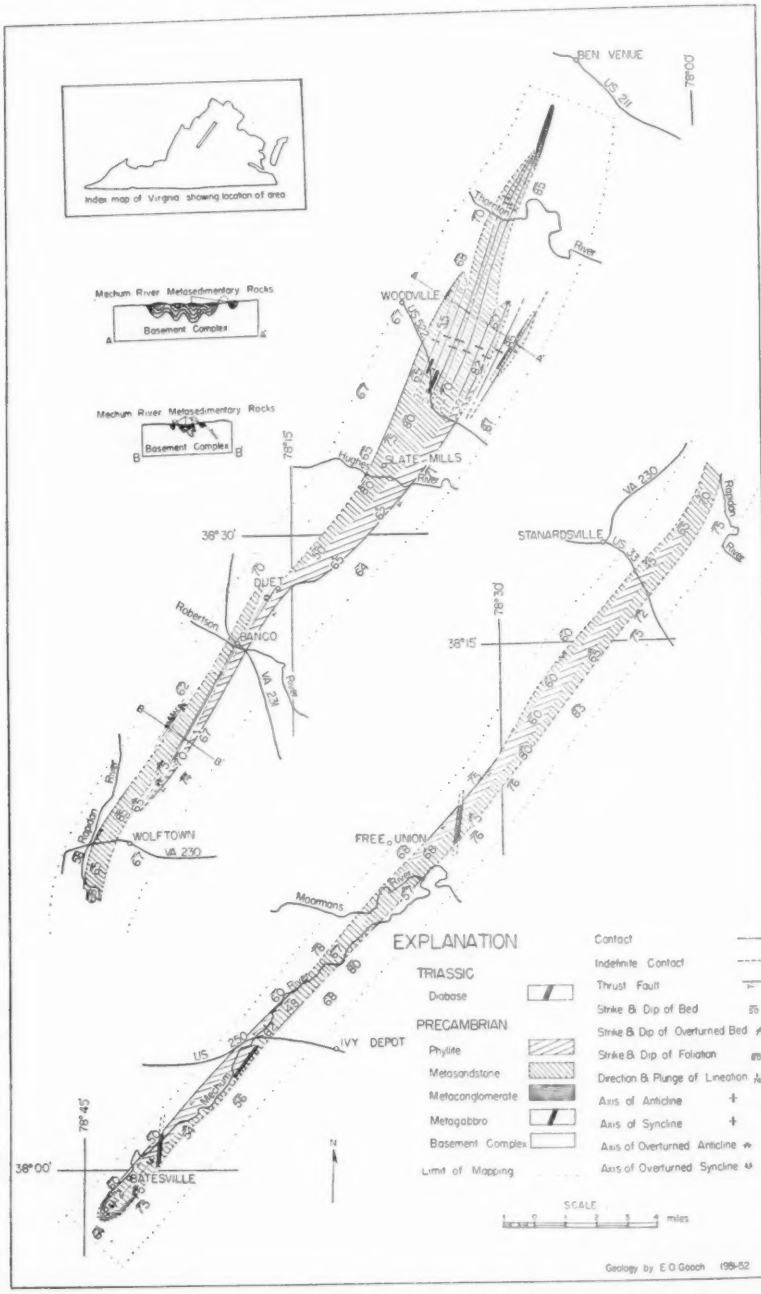

FIGURE 1.—GEOLOGIC MAP AND GENERALIZED CROSS SECTIONS OF MECHUM RIVER METASEDIMENTARY ROCKS

individuals, the writer is greatly indebted for assistance and encouragement in the preparation of this report.

formations in the Piedmont, and the northwestern limb contains the Swift Run, Catoctin, Unicoi, and higher Cambrian formations in the Blue Ridge. Between the two limbs the basement complex is exposed across a span of about 20 miles.

TABLE 1.—FORMATIONS IN THE BLUE RIDGE IN CENTRAL VIRGINIA*

Formation	Description
Unicoi Formation	About 500 feet of volcanic rocks, conglomeratic gray-wackes, subgraywackes, arkosic rocks, and quartzites.
Catoctin Formation	An undetermined thickness of greenstone interbedded with graywackes, tuffs, and arkoses. Some graywackes and arkoses are conglomeratic and have well-rounded pebbles of quartz and granite.
Swift Run Formation	For the most part 0-400 feet of graywackes, subgraywackes and volcanic rocks. It unconformably overlies the basement complex and conformably underlies the Catoctin Formation. The volcanic rocks include andesites, tuffs, and greenstones. Considered by Bloomer (1950, pp. 760–775) to be equivalent to uppermost part of Lynchburg.
Lynchburg Formation	An unknown thickness of conglomeratic gneiss, biotite gneiss, pebble quartzite, and phyllite. The conglomeratic gneiss, known as the Rockfish, occurs at base of Lynchburg in lenticular bodies. Biotite gneiss occurs in beds from a few inches to many feet in thickness and is most abundant in lower part of formation. Pebbly quartzite forms thick beds separated by beds of phyllite and gneiss and is most abundant in middle of formation. The phyllite is most abundant in upper part of formation but is interbedded with biotite gneiss and pebbly quartzite.
Basement complex	A series of gneissoid and granitoid rocks known as the Lovingston, Marshall, and Pedlar formations. The Lovingston is a gneissic and granitic formation consisting of quartz, potassium feldspar, plagioclase feldspar, and biotite. The Marshall Formation comprises the major part of the basement complex. It is a medium-grained gneiss consisting of quartz, potassium feldspar, plagioclase feldspar, and quartz. The Pedlar Formation is an assemblage of granitic, granodioritic, syenitic, quartz-dioritic, anorthositic, and unakitic rocks.

*Table compiled from Bloomer and Werner, 1955; Bloomer, 1950; Stose and Stose, 1946.

This paper is abridged from a Ph.D. thesis submitted to the Graduate Faculty of the University of North Carolina; the interest and assistance of the staff of the Department of Geology is greatly appreciated.

REGIONAL GEOLOGY

In central and northern Virginia a Precambrian basement complex of gneisses and granitoid rocks is overlain unconformably by Late Precambrian and Early Cambrian metasedimentary and volcanic formations (Table 1). The gneisses and granitoid rocks occur in the core, and the metasedimentary and volcanic formations occur on the limbs of the Catoctin Mountain-Blue Ridge anticlinorium. The southeastern limb of this structure contains the Lynchburg, Catoctin, and possibly other

INFOLDED METASEDIMENTARY ROCKS

General

The rocks in the infolded belt include metamorphosed conglomerates, sandstones, and siltstones that are intruded by metagabbro and diabase. Because of the size of the infold and the geologic significance of the metasedimentary rocks, the writer proposes that the infold be named the Batesville syncline for the village of Batesville located within the infold near its southern end and that the metasedimentary rocks be given a separate name, Mechum River metasedimentary rocks, for their exposures along Mechum River.

Conglomerates

Conglomerates of Rockfish type occur at scattered localities along both edges of the infold. Best exposures are found at its south and north ends; at the south, a conglomerate wraps around the end and attains a thickness of perhaps 500 feet (Pl. 1). At the north, conglomerate crops out along the strike for more than 2 miles and is about 200 feet thick. Other small, isolated outcrops occur along the edges of the large infolds (Fig. 1), as well as along the western edge of the small infold east of Woodville where the conglomerate is several hundred feet thick (Pl. 1).

The conglomerates are made up of boulders as much as 2 feet in diameter, derived from the various facies of the basement rocks on which they lie. Most of the boulders are granitoid, a few are gneissic, and they are composed for the most part of quartz, potassium feldspar, oligoclase, and biotite. The matrix has a similar mineral composition and consists largely of quartz, biotite, and potassium feldspar, and some plagioclase feldspar, probably oligoclase. The conglomerates grade into the overlying metasandstones, but where they are absent the latter lie directly on the basement rocks.

Metasandstones

Metasandstones make up the greater part of the Mechum River metasedimentary rocks. Where the conglomerates are present, the metasandstones grade downward into the conglomerates, but where they are absent, the metasandstones rest directly on basement rocks. Most of the beds vary in thickness from several inches up to about 2 feet, although a few are massive. Bedding is well preserved in the thinner beds (Pl. 2), and most of these show graded bedding. Thin beds of phyllite are interbedded with the metasandstones.

The metasandstones, derived from the nearby basement rocks, vary in composition according to the nature of the source rocks. They consist for the most part of quartz, potassium feldspar, some plagioclase, and considerable biotite in places, especially in the southern part of the infold. Accessory minerals include epidote, ilmenite, apatite, and pyrite. Percentage of any one mineral is variable along the strike. Just south of Thornton River the original rock was a fairly pure quartz sand, whereas along the western edge of the infold near Banco it was an arkose. In other places, as along Mechum River south of Free Union, it was a graywacke in that it was poorly sorted with considerable matrix material.

Phyllites

In addition to being interbedded with the metasandstones as thin beds, phyllites crop out in two large areas in the infold in the southern part of the infold and along the eastern edge of the infold to the north (Pl. 2). They are quartz, sericite, biotite, and oligoclase, and some contain minor amounts of graphite. In some of the phyllites there is an A-lineation expressed by mineral streaks and grooves. The phyllites are associated with zones of reverse faulting, some of which has destroyed bedding; in areas of less movement, bedding has been distorted but is still present.

Basic Dikes and Sills

Basic dikes and sills of two geologic ages intrude the metasedimentary rocks. The older intrusive rocks are metagabbros which are scattered throughout the infold and strike northeastward-southwestward, parallel to the strike of the metasedimentary rocks; they may be sills. The youngest are diabase dikes of Triassic age which strike southward. Two of these intrude the metasedimentary rocks in the southern part of the infold.

Structure

The Mechum River metasedimentary rocks form an infold near the axis of the Catoctin Mountain-Blue Ridge anticlinorium (Fig. 2), separated from the limbs of the anticlinorium by broad expanses of basement rocks. The infold consists of two large echelon synclines, one extending from the southern tip northeastward to a point about a mile west of Duet, and the other from northeast of Wolftown northeastward to the northern tip. Between Duet and Wolftown the two folds overlap and are separated by a narrow anticline of basement rocks. Two small synclinal areas of metasedi-

PLATE 1.—CONGLOMERATE OUTCROPS

FIGURE 1.—Conglomerate along western edge of small infold east of Woodville.
FIGURE 2.—Large boulder in conglomerate at southern end of infold.

FIGURE 1

FIGURE 2

CONGLOMERATE OUTCROPS

FIGURE 1

FIGURE 2

PHYLLITE AND METASANDSTONE OUTCROPS

mentary rocks lie outside the main belt of Mechum River metasedimentary rocks and are separated from it by anticlines of basement rocks; the larger is east of Woodville and the other north of Wolftown.

graded bedding, which is especially well shown along the southeastern edge of the infold along Moormans River. To the north the folds become open, especially from Slate Mills northward (Section *AA'*, Fig. 1).

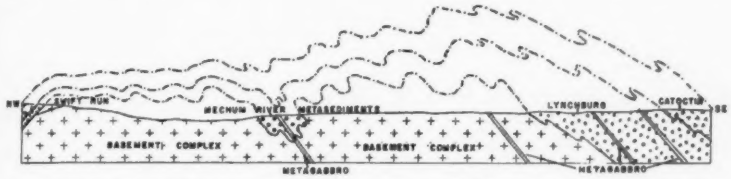

FIGURE 2.—GEOLOGIC CROSS SECTION OF PART OF CATOCTIN MOUNTAIN-BLUE RIDGE ANTICLINORIUM
Modified after Bloomer, 1950.

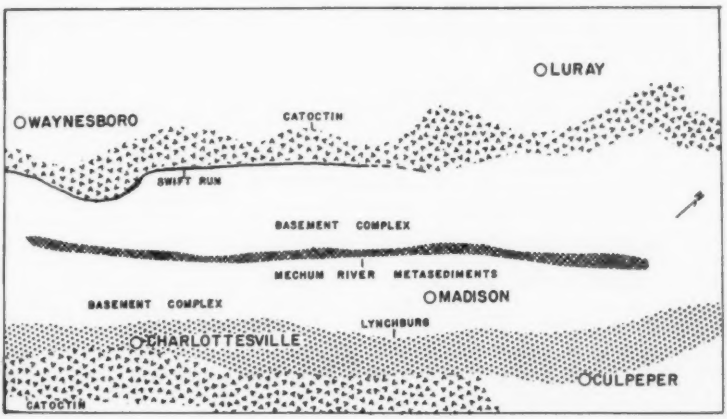

FIGURE 3.—INDEX MAP OF NORTH-CENTRAL VIRGINIA
Geographic relation of Mechum River metasedimentary rocks to Lynchburg and Swift Run formations is shown.

Although the contact between basement rocks and metasedimentary rocks is faulted in places the metasedimentary rocks have escaped erosion mainly because of downfolding. Best evidence for the synclinal structure is the manner in which the infold pinches out to the north, where a series of synclinal "fingers" extend into the basement rocks. Additional evidence is the basal conglomerates along the edges and at both ends.

The southern part of the infold is very compressed, and the beds on the southeastern limb are overturned northwestward (Section *BB'*, Fig. 1). Overturning of the beds is proved by

Faults

Reverse faults locally form the contact between basement rocks and the Mechum River metasedimentary rocks; but the writer found no evidence that any of them have great displacement. They are part of a distributive system, lesser members of which occur on each side, within the metasedimentary rocks and in the basement rocks. Such faults were observed along the western edge of the infold at the crossing of U. S. Highway 250 over Mechum River where foliation is as well developed in

PLATE 2.—PHYLLITE AND METASANDSTONE OUTCROPS

FIGURE 1.—Phyllite along eastern edge of infold near Slate Mills.
FIGURE 2.—Bedding in metasandstone along Mechum River.

the basement rocks as in the metasedimentary rocks.

Most of the faulting occurs in the southern part of the infold where it is present on both flanks. No evidence of faulting was found north of U. S. Highway 522.

Overturning of the folds, reverse faulting, and width of the metasedimentary rocks are directly related. In the northern part of the infold the folds are open, faulting is absent and the metasedimentary rocks attain their maximum width. Southward along the strike the belt of metasedimentary rocks narrows, faults develop, and folds become overturned.

Stratigraphic Relationships

The Mechum River metasedimentary rocks lie about halfway between the Swift Run Formation in the Blue Ridge to the west and the Lynchburg Formation in the Piedmont to the east (Fig. 3). The Mechum River metasedimentary rocks, the Lynchburg Formation and the Swift Run Formation were all derived from the basement complex and rest upon it unconformably.

Bloomer (1950, p. 753–783) interprets the Lynchburg Formation as overlapping the basement complex from east to west. The Oronoco Formation (Bloomer and Bloomer, 1947, p. 94–106) and the Swift Run Formation (Stose and Stose, 1946, p. 11–131) represent the thinned western edge of this overlap.

Brown (unpub. manuscript) reports that near Lynchburg, Virginia, pebbles in a conglomerate at the base of the Lynchburg came from the basement complex to the west.

The writer believes that the Mechum River metasedimentary rocks are equivalent, in part, to the Lynchburg and Swift Run formations.

Evidence for Correlation with the Lynchburg and Swift Run Formations

The east to west overlap of the Lynchburg and Swift Run formations on the basement complex suggests that the Mechum River metasedimentary rocks, located about halfway between, are at least in part equivalent to them. The Mechum River metasedimentary rocks were derived from the basement complex and overlie it unconformably just as do the Lynchburg and Swift Run formations.

The Catoctin Formation overlies the Lynchburg and Swift Run formations in central Virginia. The Mechum River metasedimentary rocks were also probably overlain by the Catoctin, but it has been removed by erosion.

The basement complex, the Lynchburg and Swift Run formations, and the Mechum River metasedimentary rocks are intruded by basic dikes and possibly sills, mostly metagabbros, but none occur in formations above the Catoctin greenstone. Occurrence of the dikes and sills in the Mechum River metasedimentary rocks strengthens the conclusion that the latter are equivalent to the Lynchburg and Swift Run.

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DRAINAGE EVOLUTION IN NORTHEASTERN MONTANA AND NORTHWESTERN NORTH DAKOTA

BY ARTHUR DAVID HOWARD

ABSTRACT

Topography and the distribution of gravels support suggestions by earlier investigators that the master streams of northeastern Montana and northwestern North Dakota flowed northward across the present trench of the Missouri until diverted in the Pleistocene. The age of the prediversion drainage is not precisely known. The ancestral streams cross early Tertiary structures. It has been suggested that the streams are antecedent and are pre-Tertiary. However, middle or later Tertiary superposition from a high-level erosion surface seems equally possible.

Miocene-Pliocene (?) gravels along the ancestral valleys indicate that the streams date back at least to the latter part of the Tertiary. A high-level gravel of possibly Oligocene or Miocene age parallels Yellowstone Valley for long distances, suggesting that the Yellowstone may have followed approximately its present path across the area in mid-Tertiary time.

The ancestral Missouri, Yellowstone, and Little Missouri rivers were diverted to form the present Missouri probably in Kansan time. This is earlier than the date proposed for the glacial Missouri in South Dakota, but the dates are not necessarily contradictory. Little Missouri River was diverted again by early Wisconsin ice and joins the Missouri 80 miles farther downstream than it did formerly.

Both the Yellowstone and Missouri hug the right sides of their valleys and apparently did so in pre-Pleistocene time. Uptilting to the northwest or the effects of terrestrial rotation are possible explanations. Further work is required to test these possibilities.

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INTRODUCTION AND ACKNOWLEDGMENTS

This report is a by-product of a larger study aimed at a regional interpretation of the Cenozoic history of northeastern Montana and northwestern North Dakota with emphasis on the Pleistocene. The results of the larger study are in press with the U. S. Geological Survey. The present report draws heavily on the observations and conclusions of prior investigators, many of whom are listed in the bibliography. Other authors have contributed through observations incorporated in numerous quadrangle and area maps not listed in the

references. Among these are William E. Benson, Roger R. Colton, Garland B. Gott, Fred S. Jensen, Clifford A. Kaye, David R. Larrabee, Richard W. Lemke, Robert M. Lindvall, Chilton E. Prouty, J. Hiram Smith, Roland C. Townsend, and Irving J. Witkind. The author is indebted to Eliot Blackwelder, Stanley Davis, C. Max Bauer, and W. M. Laird for critically reading the manuscript. Only the writer, however, is responsible for conclusions reached.

Physiographic Setting

The area of investigation comprises about 20,000 square miles of the Great Plains in northeastern Montana and northwestern North Dakota (Fig. 1). The plains surface, designated the Missouri Plateau peneplane (Howard, in press), is degradational, bevelling locally deformed early Tertiary and Cretaceous sedimentary rocks. It was called the Number 2 bench by Alden (1932, p. 32), who applied the term bench indiscriminately to the tops of mesas or plateaus and to typical terraces. The term peneplane, as herein used, implies a near-end-stage surface of degradation; it has no climatic implications. The surface herein considered was almost certainly formed under semiarid conditions.

The plateau terminates eastward at the Missouri escarpment, a scarp several hundred feet high, which overlooks the Central Lowland. A small part of the lowland is shown in the northeastern corner of Figure 1.

Major valleys are trenched as much as 600 feet below the peneplane, and buttes, mesas, and plateaus rise about the same distance above it. The buttes and mesas in North Dakota are relatively small; the largest, Kildeer Mountains, occupy only a few square miles. The Flaxville plateaus in Montana, however, cover several hundred square miles and stand 300–600 feet above their surroundings. Where the peneplane is constricted between plateau remnants, it appears as a broad valley; where the valley is trenched, the peneplane is represented only by the broad Cartwright bench (Howard, in press). The most extensive Flaxville plateaus are north of the Missouri River. Those nearest the river form a high bench paralleling the valley. A similar bench parallels Yellowstone River and terminates westward at a high, narrow gravel-capped divide.

The divide between Atlantic and Arctic drainage wanders through the Coteau morainal belt. Most of the drainage of the region is tributary to the Missouri and flows to the Gulf of Mexico; the drainage of the northeastern lowland is to Hudson Bay. The master streams of the region are the Missouri, Yellowstone, and Little Missouri rivers. These flow in trenchlike valleys 300–600 feet deep.

Historical Review

The present path of the Missouri River is anomalous in that it crosses the region almost at right angles to the northeasterly regional slope. General G. K. Warren (1869, p. 311) suggested that the present path was determined by glacial deflection of an earlier upland drainage. In 1884 (p. 392) and again in 1914 (p. 265) and 1923 (p. 489–490), Todd suggested that the major streams of eastern Montana and western North Dakota formerly flowed northeastward into Hudson Bay. Beekly (1910, p. 323) believed that the broad swale extending northeastward from Medicine Lake (Fig. 1) represented either the preglacial valley of the Missouri or of a large tributary. Bauer (1915, p. 53) suggested that the ancestral course of the Missouri diverged from the present trench just east of Poplar, where the floor of a broad upland swale hangs about 250 feet above present river level. Bauer traced this swale into that described by Beekly northeast of Medicine Lake and proposed that together they marked the ancestral path of the Missouri into Arctic drainage. Alden (1932, p. 58) and Witkind (in press) support this view. The only objection to the proposal that the Missouri River formerly flowed northeastward into Canada before being glacially diverted to its present course came from Leonard (1916, p. 295), who maintained that the present path of the river antedated the earliest glaciation.

Alden (1932, p. 58) suggested that the ancestral upland path of the Yellowstone River coincided with the present path of the Missouri between the mouth of the Yellowstone and Williston, and that it continued north along the broad upland swale within which Little Muddy Creek has opened out its present valley (Fig. 1). Alden projected this path to Crosby through a broad swale in the Coteau morainal belt. There is no surface indication of the ancestral path beyond the Coteau morainal belt, although drilling for water has revealed a buried valley at Crosby (Waring and LaRocque, 1949, Pl. 6). Alpha (1935, unpub. M. A. thesis, Univ. of North Dakota) suggested that the Yellowstone turned back sharply to the southwest near Crosby and re-entered the plateau to join the ancestral Missouri in northeastern

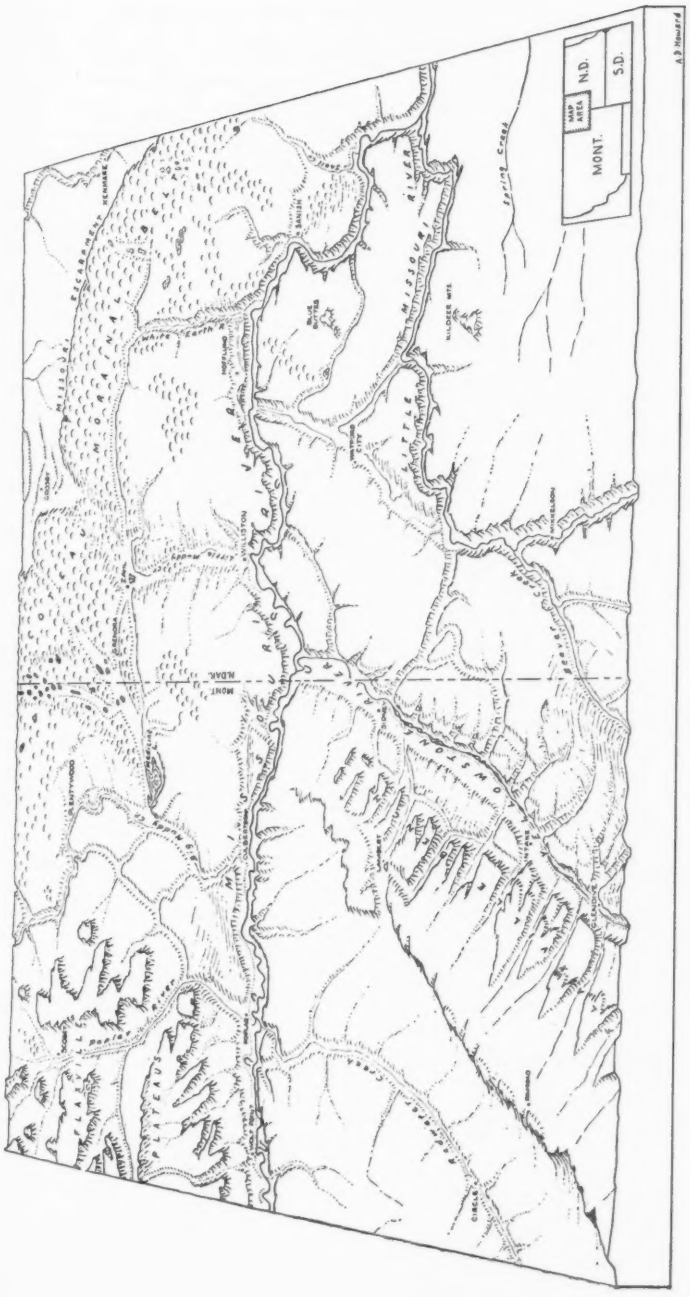

FIGURE 1.—PHYSIOGRAPHIC DIAGRAM OF NORTHEASTERN MONTANA AND NORTHWESTERN NORTH DAKOTA

Montana. This seems unlikely because of the absence of a second swale of suitable size and the lack of a reasonable explanation for the river re-entering the plateau after having once left it. Subsurface data indicate the presence of deep bedrock valleys in southern Saskatchewan, which may be the preglacial courses of the Missouri and Yellowstone rivers (Meneley *et al.*, 1957 (p. 444–445).

Little Missouri River, in the southeastern part of the area (Fig. 1), turns abruptly from a northerly path to follow an irregular path to the east. Between the abrupt turn and the Missouri River 50 miles to the northeast is a continuous upland swale, the floor of which is, in places, up to nearly 2 miles wide and is occupied by several independent streams. Attention was first directed to this swale by Wilder (1905, p. 43), and the feature was described in detail by Leonard (1916, p. 300–301). Edwards and Ableiter (1942, p. 5–6) found that the alluvium on the floor of the swale continued without break from one end to the other and agreed with Wilder and Leonard that the swale probably represented the ancestral course of Little Missouri River. They suggested, however, that this was the preglacial course, whereas evidence to be presented in this paper indicates that the swale was probably occupied until early Wisconsin time. The upland swale appears to cross the Missouri just west of Hofflund and to continue east to White Earth Valley, with which it coincides for at least several miles to the north.

Most conjecture on the drainage history of the region starts with the drainage of the peneplane. Bauer (1915, p. 56), however, suggested that the original paths of the Missouri and Yellowstone across the Great Plains and on to Hudson Bay date back to the end of the Cretaceous, when these rivers extended their courses across the plain exposed by withdrawal of the Cretaceous sea. He attributed deposition of the early Tertiary sediments of the plains to these rivers and their neighbors. He believed that the master streams were antecedent with respect to early Tertiary domes and anticlines. This is possible, but it seems equally possible that the rivers were superimposed on these structures from a higher surface, perhaps from the Oligocene or Miocene (?) surface referred to later. Superimposition would necessarily have taken place at a level higher than the peneplane, because the Missouri and Yellowstone were already established in approximately their present paths in Flaxville (Miocene-Pliocene ?) time, the cycle preceding the one in which the peneplane was developed. This is indicated by the high-level gravel-capped Flaxville benches along these valleys.

Although the early Tertiary drainage of the region is for the most part a matter of speculation, there is evidence of the presence of large streams in this area in middle and late Tertiary time. Alden (1932, p. 12) and Collier and Knechtel (1939, p. 13) described patches of coarse, fluvial gravel on the divide west of Yellowstone Valley, 1200–1300 feet above Yellowstone River. Howard has named this the Rimroad gravel (in press). Alden correlated this gravel with previously described and similarly situated gravels elsewhere and dated it as Oligocene or Miocene. He suggested that the high divide was once a valley bottom. Mackin (1937, p. 868–873) has questioned the assigned age of the high-level gravels with which the Rimroad gravel has been correlated and suggests that the former may be younger than heretofore supposed. Because of this possibility, the dates assigned in this article should be regarded as tentative only.

Between the Rimroad divide and Yellowstone River is a broad bench capped by Flaxville gravel. Similar gravels cap a bench along Missouri River. These benches, as well as scattered plateau remnants, stand 300–600 feet above the Missouri Plateau peneplane. On the basis of vertebrate fossils in the gravel they have been dated as upper Miocene or lower Pliocene (?). The Flaxville gravel thus records a second high level of fluvial activity prior to the development of the Missouri Plateau peneplane.

A third gravel, named the Cartwright gravel by Howard (in press), marks the paths of streams across the peneplane itself. The gravel caps the Cartwright bench along parts of the major valleys.

Cumulative Evidence For Northeasterly Ancestral Drainage

Types of Evidence

The evidence favoring a northeasterly ancestral drainage is in small part factual and in large part inferential. It consists of: (1) upland swales; (2) parallelism of upland swales and Flaxville benches; (3) fluvial gravel in upland swales; (4) fluvial pebbles in till northeast of known gravel exposures; (5) variations in width of Missouri Valley; and (6) barbed tributaries, wind gaps, and acute bends in the Missouri and Little Missouri valleys.

Upland Swales

Broad swales (Fig. 1) diverging from the valleys of the master streams have dimensions comparable to those of the present valleys at the level of the Cartwright bench. The floor of the upland swale trending northeast from Poplar stands about 250 feet above the floor of Missouri Valley, as does the Cartwright bench west of Poplar. The width of the swale, measured from the foot of the Flaxville plateaus on the northwest to the drainage divide on the southeast, is about 12 miles. This agrees closely with the width of the Missouri Valley at the Cartwright level.

Similarly, the upland swale north of Williston and its continuation through the Coteau morainal belt south of Crosby (Fig. 2) are aligned with Yellowstone Valley and have the same width (15 miles), from crest to crest, as does Yellowstone Valley at the level of the Cartwright bench (Fig. 3).

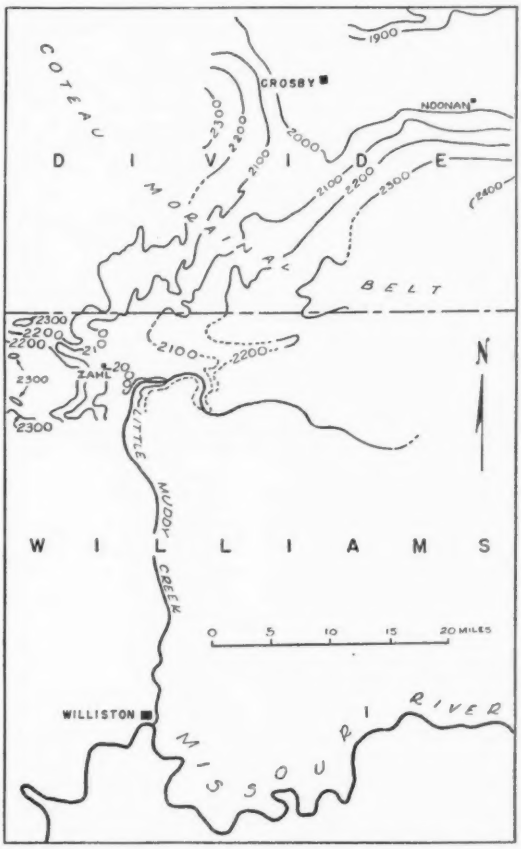

FIGURE 2.—SWALE THROUGH COTEAU MORAINAL BELT

This swale, in northwestern North Dakota, probably represents the partially buried ancestral valley of Yellowstone River.

The upland swale between Little Missouri Valley and the Missouri River is 6–8 miles wide. This agrees reasonably well with the breadth of Little Missouri Valley at the same level to the south. However, the writer has examined the southern area only cursorily.

Thus, on the basis of alignment and dimensions, the upland swales may well represent the ancestral valleys of the Missouri, Yellowstone, and Little Missouri rivers.

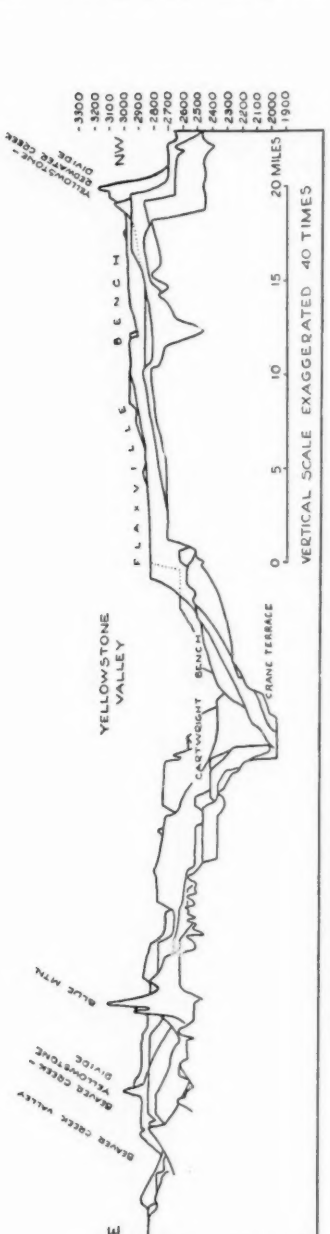

FIGURE 3.—PROJECTED PROFILES OF YELLOWSTONE VALLEY AND BORDERLANDS

The width of the block is 60 miles; the length along the valley, 40 miles. The surfaces descend toward the observer, hence appear higher in the background. The Cartwright bench is the correlative of the Missouri Plateau peneplane here constricted between the high Flaxville remnants. The principal erosion surfaces are composite, consisting of a number of shallow terraces. Vertical exaggeration 40 times.

Parallelism of Upland Swales and Flaxville Benches

The upland swale trending northeast from Poplar diverges from the Missouri Valley with the same trend as the Flaxville bench hundreds of feet higher. The parallel trends are shown in a generalized contour map of the areh (Pl. 1) constructed from available topographic maps by eliminating all but major dissection. The relationships suggest that in Miocene-Pliocene (?) time, when the Missouri River flowed at the Flaxville level, it drained to the northeast as suggested by Alden (1932, p. 21). The parallel trend of the swale hundreds of feet below suggests that the drainage continued in that direction until the development of the Missouri Plateau peneplane in early Pleistocene time.

The Flaxville bench south of the Missouri nowhere diverges from the present valley of Yellowstone River, indicating that the ancestral and modern paths, upstream from the present confluence with the Missouri, have coincided since at least Miocene-Pliocene (?) time.

Fluvial Gravel in Upland Swales

According to Alden (1932, p. 50) the Cartwright bench has a gradient of about 1.75 feet per mile for 225 miles above Poplar. The floor of the swale northeast of Poplar, however, is for the most part below the projected profile of the bench, which is attributed by both Witkind (in press) and the writer to glacial erosion by ice lobes following the swale to the south. Gravel veneered with till, however, caps a high bedrock spur southwest of Medicine Lake. The gravel is at an altitude of about 2020 feet (Witkind, Personal communication). The writer examined this gravel in the field with Witkind and found that it superficially resembles the Flaxville, Cartwright, and Crane Creek gravels. The Crane Creek gravel, which was named by the writer after a tributary of the Yellowstone River (Howard, in press), caps a broad strath terrace, the Crane terrace, within the major valleys. Witkind made a pebble count and found that the lithology of the gravel at Medicine Lake is almost identical to that of the Flaxville gravel. It consists, for the most part, of quartzites, with lesser amounts of chert and quartz. The gravel, however, is hundreds of feet lower than any known occurrence of Flaxville gravel in the vicinity. The southeasterly slope of the Flaxville bench (Pl. 1) averages less than 10 feet per mile in its flatter, distal parts. If projected the 17 miles to Medicine

Lake, the cross-profile of the bench would pass over the Medicine Lake gravel locality at an elevation of 2430 feet, more than 200 feet higher than the gravel. It seems unlikely, therefore, that the gravel is Flaxville. The absence of glacial pebbles of northeastern origin distinguishes it from the Crane Creek gravel, which contains many such pebbles. The Medicine Lake gravel is believed to be Cartwright gravel in spite of the absence of scattered glacial pebbles which appear in exposures of the Cartwright gravel elsewhere. The gravel-capped bedrock spur is probably a remnant of the floor of the ancestral valley that escaped destruction during subsequent glacial erosion.

Gravel on the floor of the upland swale northeast of Williston proves that the swale was once occupied by the Yellowstone River. The high-level gravel is on the west side of Little Muddy Valley, about 12 miles north of Williston. It resembles the Cartwright and other gravels along Yellowstone Valley in the abundance of acidic and basic volcanic rocks, types that are rare or absent in Missouri River gravels upstream from the Yellowstone confluence. The gravel was presumably transported to its present position by Yellowstone River before the ancestral valley was crosscut by the Missouri trench.

High-level gravel has been found in the postulated ancestral valley of Little Missouri River south of the Missouri trench, but not to the north. The area to the north, however, is thickly mantled by till.

Fluvial Pebbles in Till North of Known Gravel Exposures

The surface till of the region includes many of the distinctive brown, water-worn pebbles that characterize the Flaxville, Cartwright, and Crane Creek gravels. The concentration is greatest in and leeward (southwest) of large areas of exposure of the gravels. The northeasternmost gravel exposures found to date are in northwestern North Dakota (Howard, Gott, and Lindvall, 1946). Distinctive pebbles, however, are found in till still farther north and east (Howard 1956, p. 1205, Fig. 5) suggesting that gravel is present under the till northeast of known exposures. Although the buried gravels could have been deposited by streams other than those herein considered, the proximity to the postulated ancestral courses of the Missouri and Yellowstone suggests that these streams were probably responsible for their deposition.

Variations in Width of Missouri Valley

Calhoun (1906, p. 34–45) described abrupt variations in the width of Missouri Valley in central Montana. He found that the broad segments corresponded to places where the post-glacial river reoccupied parts of its own or other drift-filled valleys and that the narrow reaches were new paths acquired by glacial diversion. This situation is remarkably well shown in northeastern Montana and northwestern North Dakota. Here, the Missouri Valley is broad in precisely those places where it coincides with postulated ancestral courses, and narrow in those segments which cross postulated ancestral divides (Fig. 4).

From the west edge of Figure 4 to a point just east of Poplar, the width of the Missouri Valley below the level of the Cartwright bench (peneplane level) is 3½–4 miles. The presence of the Cartwright bench indicates that the present and high-level courses coincide in this stretch of valley. Between this point and the mouth of the Yellowstone the trench cuts across a northeasterly trending belt of high country. The trench narrows to 1–2 miles, and the Cartwright bench is missing. Between the mouth of the Yellowstone and Williston the valley abruptly widens again to 3½–4 miles, and the Cartwright bench is present at least locally. This broad part of the Missouri Valley is in line with Yellowstone Valley and the upland swale north of Williston. Inasmuch as the semiconsolidated sedimentary formations in the area are more or less uniformly resistant to weathering and erosion, the narrowness of the trench between Poplar and the mouth of the Yellowstone is reasonably explained on the assumption that it is younger. According to this interpretation, after the river reached grade, it was unable to widen its valley as rapidly in this high area as elsewhere because of the greater thickness of material to be removed.

East of Williston, the trench narrows again, but there is no broad segment at the confluence of the high-level ancestral valley of Little Missouri River. This is not entirely unexpected, however, inasmuch as the present valley of the Missouri cuts across rather than coincides with the postulated ancestral path of the Little Missouri.

The narrowest parts of the present trench of the Missouri, southeast of Culbertson and south of Sanish, are at sites where evidence indicates local displacements of the river subsequent to the major diversion.

Thus, the distribution of the broad and narrow segments in the Missouri Valley are reasonably explained on the assumption that the ancestral drainage was to the northeast and that barbed tributaries are in alignment with tributary valleys on the north side of the master streams, suggesting bisection of ancestral tributaries by the present trenches.

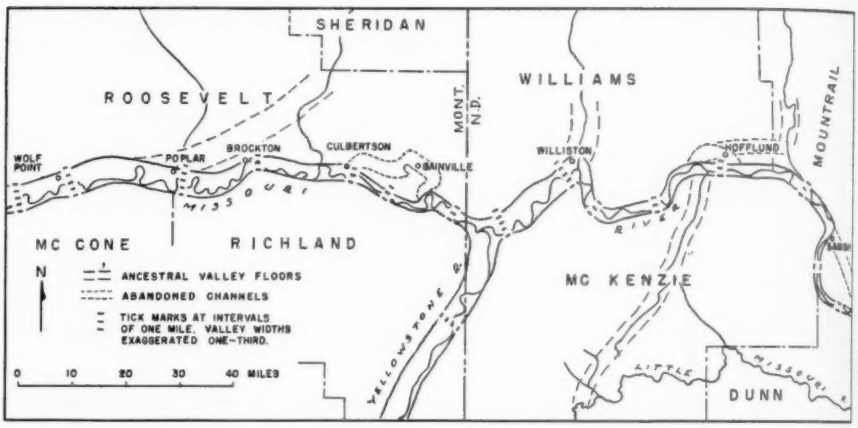

FIGURE 4.—VARIATIONS IN WIDTH OF MISSOURI VALLEY BELOW LEVEL OF MISSOURI PLATEAU PENEPLANE

Broad segments coincide with ancestral valleys; narrow segments cross ancestral divides. The principal ancestral valleys are shown by long dashes; abandoned channels, by short dashes.

the present narrow segments were eroded later across the ancestral divides.

Barbed Tributaries, Wind Gaps, and Acute Bends in Missouri and Little Missouri Valleys

A number of tributaries of the Missouri and Little Missouri rivers enter the present trenches with barbed relationships. Several of the larger barbed tributaries are shown in Figure 5, as, for example, Charley and Hardscrabble creeks between Poplar and Culbertson, an unnamed creek southeast of Williston, and Bennett and Bear creeks tributary to the Little Missouri. These same creeks, however, are not barbed with respect to the ancestral valleys. A wind gap at Brockton, on the narrow divide between the present and ancestral valleys of the Missouri, is in line with Charley Creek and may represent the former continuation of the latter into the ancestral valley. A wind gap on the opposite side of the ancestral valley, where Poplar River turns sharply to the southwest, probably marks the former point of entry of Poplar River into the ancestral valley (Alden, 1932, p. 44).

Hardscrabble, Bennett, Bear, and other Some barbed tributaries, such as Bear Creek and the one southeast of Williston, are aligned with unusual, acute bends in the major valleys, suggesting that the course of the present trenches may have been controlled in part by the orientation of tributaries to the ancestral valleys.

In brief, these peculiarities of the present drainage present no problem if one assumes that the secondary streams were originally tributary to the postulated ancestral valleys and that the present trenches were eroded later.

EVIDENCE FOR GLACIAL DIVERSION OF ANCESTRAL DRAINAGE

Missouri Valley, within this area, is one of three deep east-west trenches of great linear extent (Fig. 5). All three are at or within the glacial limit; none are found to the south. This suggests an origin related to the glacial history of the region.

The southernmost trench, at the glacial limit, coincides with the border of the early Wisconsin drift. The channel is locally 3 miles wide and more than 400 feet deep and is occupied by a number of independent streams. Except for Little Missouri River in the eastern part of the area, these streams are relatively

small. E. R. Schmitz (1955, unpub. M.A. thesis, Univ. of North Dakota) suggested a sequence of captures to explain the easterly path of the pendent streams. In the western part of the area it follows the border of the late Wisconsin drift closely; elsewhere it diverges appreciably.

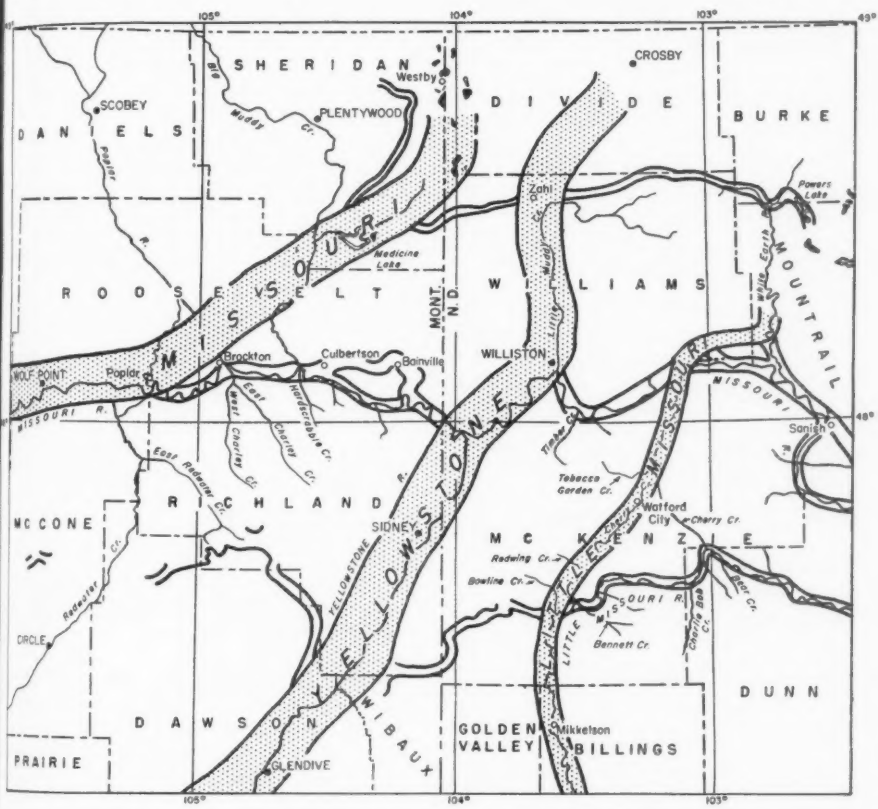

FIGURE 5.—ANCESTRAL DRAINAGE AND REGIONAL EAST-WEST ICE-MARGINAL CHANNELS

The present trench of the Missouri is one of three east-west channels cutting across the ancestral paths of the master drainage. A number of tributaries to the ancestral streams are shown. Several of these may have influenced the paths of ice-marginal waters. The widths of the ancestral valleys are approximate valley floor widths.

Little Missouri River. Available evidence fails to support this hypothesis. The channel is almost certainly an ice-marginal feature. Little Missouri River apparently maintained its easterly diversion path after recession of the early Wisconsin ice, whereas Yellowstone River resumed flow northward to the Missouri.

A second east-west channel crosses the northern part of the area between Medicine and Powers lakes. Except that it is shallower and narrower, it resembles the channel at the glacial limit. It too is occupied by a number of independent streams. In the western part of the area it follows the border of the late Wisconsin drift closely; elsewhere it diverges appreciably.

The present trench of the Missouri is thus one of three large east-west channels. It differs principally in that it is occupied throughout by a major stream; the southern trench is occupied by a major stream only in its eastern part; the northern trench is nowhere occupied by a major stream. Analogy with the southernmost channel supports the view that the present trench of the Missouri is an ice-marginal channel. Damming of north-draining valleys presumably created lakes which spilled across and trenched the intervening divides.

Date of Diversion

The early Wisconsin till,[1] which extends to the glacial limit, is present within the trench of the Missouri. The till not only mantles the Crane Creek gravel but overlaps the edge of the Crane terrace and extends an unknown distance below the valley floor. The Crane Creek gravel is dominantly nonglacial; the terrace it mantles marks the culmination of a post-diversion interglacial erosion cycle. At the close of this cycle the floor of the diversion trench of the Missouri was several hundred feet below the upland and about 4 miles wide. A second significant erosion cycle, the Deep-trench cycle, is recorded by a deep inner trench eroded as much as 200 feet below the Crane terrace, as revealed by drilling (Fig. 6). Inasmuch as the early Wisconsin till drapes over the interglacial Crane terrace and extends down into the inner trench, it is later than both these erosion cycles, or, conversely, the diversion trench antedates the early Wisconsin by at least two erosion cycles.

Other evidence, beyond that suggesting that the present Missouri trench is an ice-marginal channel, indicates that the diversion took place in Pleistocene time. The ancestral upland valleys of the Missouri and Yellowstone rivers are peneplane valleys; together with the broad interfluves they form the Missouri Plateau peneplane. There is no indication that the ancestral rivers continued to occupy these valleys at a lower level, whereas the presence of Crane Creek gravel within the present diversion trench indicates that the trench was excavated during the first post-peneplane cycle of erosion. Thus, the diversion is post-peneplane and pre-Wisconsin. The age of the peneplane is thus important in dating the diversion. Available evidence suggests that the peneplane is Pleistocene. Belts and patches of Cartwright gravel on the peneplane mark either higher levels of present streams or the paths of former streams. Although no fossils were found in these gravels by the writer, fragments of bone and a single tooth were found by Collier and Thom (1918, p. 182) in possibly equivalent gravels west of the present area of investigation. The tooth was identified by Doctor Gidley of the U. S. National Museum as that of a horse resembling the living species. Doctor Gidley did not believe that it could be as old as the fauna of the Flaxville gravel (Miocene-Pliocene?).

Alden (1932, p. 44) also regarded the pene-

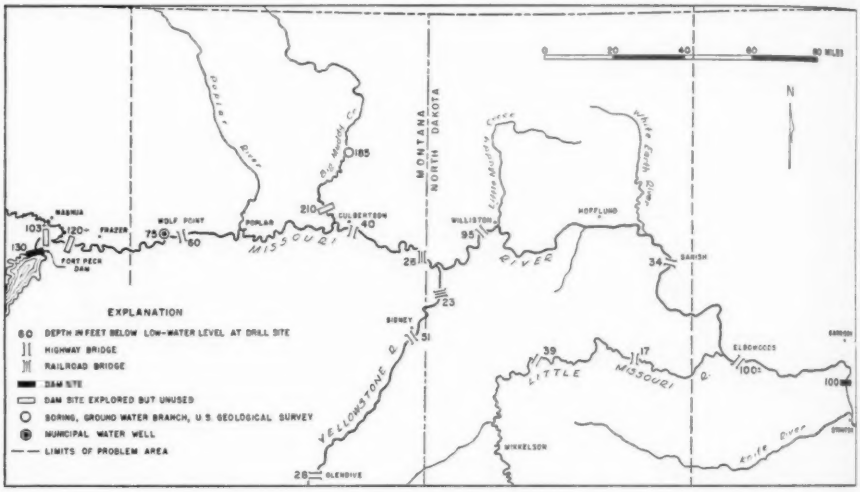

FIGURE 6.—Bedrock Depths in Major Valleys

Bridge borings are generally confined to the vicinity of the river crossing, hence do not necessarily indicate the maximum depth to bedrock within the valley. A depth of 175 feet was recorded near Fort Peck Dam. Data from U. S. Corps of Engineers, Bureau of Reclamation, State Highway Commissions, U. S. Geological Survey, Great Northern Railway, and local municipalities.

[1] The dating of this and other tills is discussed in detail by the writer in a forthcoming Professional Paper of the U. S. Geological Survey.

plane and the gravels (Cartwright) as Pleistocene. He found an old till on remnants of the Flaxville plain near the mountains and concluded that the Flaxville plain "had not been very deeply dissected before the extension of the first of the mountain glaciers onto it". He believed that if the Flaxville surface was not itself Pleistocene, then Pleistocene time began with an uplift that caused the streams to cut below the Flaxville surface. In his opinion, Pleistocene time began "a long time prior to the completion of the second set of terraces or No. 2 bench [Missouri Plateau peneplane] by planation and spreading of the gravel thereon." He regarded the peneplane as the first of the Pleistocene "benches".

That the surface in North Dakota and Montana is Pleistocene is indicated by scattered glacial pebbles in the Cartwright gravel. C. R. Warren (1952, p. 1148) found vertebrate fossils in gravel occupying a swale in what is probably the Missouri Plateau peneplane near Chamberlain, South Dakota. The fossils were identified by C. Bertrand Schultz of the Nebraska Museum, who suggested a Yarmouth age. Although the South Dakota locality is far from the present area of investigation, it seems reasonable to suppose that the surface is Pleistocene in North Dakota as well.

That the drainage diversion which formed the Missouri River in North Dakota and northeastern Montana did not take place as long ago as Nebraskan time is suggested by the relatively narrow belts of dissection that have developed along the diversion trench and the narrowness of the trench where it crosses the ancestral divides. In the divide areas, the trench floor is not wide enough to accommodate fully developed meanders even at the surface of the valley fill. More probably, the diversion took place either in Kansan time or, as C. R. Warren (1952, p. 1152–1154) suggests for South Dakota, in Illinoian time. An Illinoian date for the diversion in the North Dakota-Montana area seems unlikely because of the necessity of accounting for two lengthy cycles of erosion (Crane and Deep-trench) and a probable intermediate glacial advance between the date of the diversion and the beginning of the Wisconsin. These difficulties are circumvented by assuming that the diversion took place in Kansan time.

The Kansan date for the diversion in the North Dakota-Montana area does not necessarily conflict with Warren's Illinoian date in South Dakota. The diversion need not have taken place simultaneously throughout the northern Great Plains. If, however, the diversion did take place earlier in the northern area, the newly formed Missouri must have connected with the south-flowing master stream in James River Valley east of the Missouri trench in South Dakota. Future investigation, by proving or disproving the existence of such a connecting link, may help to settle this question.

At present, although there is room for disagreement, the author favors the view that the ancestral Missouri, Yellowstone, and Little Missouri rivers in northwestern North Dakota and northeastern Montana were glacially diverted to form the present Missouri in Kansan time. At that time, the Little Missouri River flowed across the upland to enter the Missouri near Hofflund and continued to do so during the succeeding erosion cycle. Two broad, gravel-capped terraces are present in the ancestral upland valley at its point of departure from the present trench of Little Missouri River in southern McKenzie County. The higher terrace, which apparently represents the floor of the ancestral valley at the peneplane level, is presumably the Cartwright terrace; the lower, the Crane terrace. The Crane terrace is also exposed in the ancestral valley at its junction with the Missouri. The Crane Creek gravel descends toward the north, and imbrication indicates flow in that direction. The coarse texture of the gravel suggests deposition by a far more vigorous stream than the present small, independent creeks which handle silty alluvium almost exclusively. The evidence thus indicates that Little Missouri River was diverted a second time 50 miles south of the original diversion at Missouri Valley. Inasmuch as the southern diversion channel coincides with the border of the early Wisconsin drift, the diversion presumably took place in early Wisconsin time.

Minor displacements of the Missouri took place within its present trench. The abandoned channels at Culbertson and Hofflund are floored with early Wisconsin till. The early Wisconsin ice presumably shunted the river to its new positions. The abandoned channel southeast of Sanish, however, is floored by late Wisconsin till, and the trench to the west is less than a mile wide. The displacement here was probably effected by the late Wisconsin ice.

Summary of Possible Drainage History

There is no factual evidence bearing on the drainage of this area prior to Oligocene or Miocene (?) time. In Oligocene or Miocene (?)

time, however, a river presumably flowed northeastward across the area in a path immediately to the west of, and 1200–1300 feet higher than, the present floor of Yellowstone Valley. This is indicated by the scattered patches of Rimroad gravel on the high, western divide. Remnants of a divide overlook the Flaxville bench north of the Missouri in the western part of the area (Fig. 1), but reconnaissance failed to reveal a gravel cap.

Rejuvenation, resulting from uplift, climatic change, or both, resulted in a reversal of topography in Miocene-Pliocene (?) time. Former valley floors, protected by pervious gravels which discouraged erosion, became divides, whereas unprotected divides became valleys during the Flaxville erosion cycle. New valley floors were opened out about 600 feet below the Rimroad valley bottoms. At the Flaxville level, the Yellowstone Valley may have been as much as 40 miles wide. The Flaxville surface seems to consist of several shallow terraces descending toward the center of the valley. Regrading was presumably intermittent rather than progressive. The coarseness and thickness of the gravels are clear evidence that neither the Flaxville surface nor the lower surfaces were developed as valley-side pediments by small side streams.

A second episode of rejuvenation resulted in 300–600 feet of downcutting, leaving the gravel-covered Flaxville valley floors standing as high terraces. The terrace along the Yellowstone is still preserved; its back slope is the high western divide. North of the Missouri, however, possible remnants of the earlier divide are found only in the western part of the area, and the former Flaxville surface now consists, for the most part, of isolated plateaus. Thus, throughout much of the area north of the Missouri, there has been a second complete reversal of topography.

The Flaxville bench within Yellowstone Valley indicates that the course of the Yellowstone River in Miocene-Pliocene (?) time more or less coincided with its present course. Missouri River, at the Flaxville level, paralleled its present path eastward to Poplar, but beyond there the trend of the Flaxville surface is toward the northeast (Pl. 1). As far back as miocene-Pliocene (?) time, therefore, the course of the Missouri was presumably to the northeast, and its present path east of Poplar had not yet been acquired.

Erosion below the Flaxville level resulted in the development of the Missouri Plateau peneplane. The ancestral master streams flowed northeast across this surface (Fig. 5) probably at least until Kansan time. The streams were blocked by ice (Kansan ?), and all north-draining valleys were dammed, forming a series of proglacial lakes. Lake waters spilled across intervening divides, crossing low cols and taking advantage of favorably disposed tributary valleys as well as favorably oriented segments of the major valleys. Because the trend of the tributaries was related to the orientation of the ancestral major valleys, the path taken by the eastward-escaping waters shows unusually sharp bends and many barbed tributaries. The ancestral divides were trenched below the level of the ancestral valleys so that the master streams were unable to reoccupy their ancestral courses after withdrawal of the ice.

In the ensuing interglacial stage (Yarmouth) the Missouri deepened its valley until its floor was about 200 feet below the ancestral valley floor. The river reached grade at this level and locally widened its valley to about 4 miles. This was the Crane erosion cycle during which the broad, gravel-capped Crane strath was developed.

The Crane erosion cycle may have been terminated by advance of the Illinoian ice (the Smoke Creek glacier of Witkind). Only scattered deposits attributable to this glacier remain.

The Deep-trench erosion cycle followed, presumably in the Sangamon interglacial stage. The Crane strath was incised more than 200 feet leaving it stranded as a high terrace in the major valleys.

This was the situation encountered by the early Wisconsin ice which inundated the Missouri trench and pressed on south to the maximum limit of glaciation. The ice front remained at the glacial limit for a long time, impounding large lakes in Yellowstone and Little Missouri valleys, as recognized by earlier investigators. A third large lake was impounded in Redwater Valley. Howard (in press) has named the lake in Redwater Valley "Lake Circle"; that in Yellowstone Valley, "Lake Glendive"; and that in Little Missouri Valley, "Lake Mikkelson", after towns of these names. The overflow waters of these lakes escaped eastward along the ice front, eroding a chain of deep channels across the divides. On recession of the early Wisconsin ice, Redwater and Yellowstone rivers returned to their former northward paths, but Little Missouri River retained its newly acquired ice-marginal course, which led it into the Missouri 80 miles farther downstream than formerly. The early Wisconsin ice shifted the

Missouri to new positions at two localities, east of Culbertson and west of Hofflund (Fig. 1).

Glacial advances in the middle and late Wisconsin modified the drainage only slightly. The middle Wisconsin ice reached the Missouri near Williston, but whether it dammed the river is unknown. It did, however, contribute outwash to the fill of the valley; Williston is located on a terrace built partly of this outwash. The late Wisconsin ice, which deposited the thick moraines of the Coteau morainal belt, reached the Missouri in the eastern part of the area, blocking the Missouri and diverting it from the swale southeast of Sanish to the narrow trench to the west. Some of the delicately laminated silts that contribute to the prominent fill terrace in Missouri Valley may have accumulated in the lake formed at this time. The Mankato age of the silts is suggested by the remains of a *Bison antiquus* (?) tentatively identified by C. Bertrand Schultz of the University of Nebraska.

The valley fill was trenched to below present river level since recession of the late Wisconsin ice and was subsequently partially refilled by Recent alluvium on which the Missouri now flows.

One last aspect of the drainage of the region merits consideration. Both the Missouri and Yellowstone rivers within this area hug their right banks. Several possible explanations have been considered. The first is displacement by excess deposition from one side. Floods of outwash from the north could have displaced the Missouri against its south (right) bank, but the river now flows on silty alluvium under which the outwash is largely buried. There is no reason for believing that more silt has been brought into the valley from the north than from the south. In any event, outwash-laden streams could not have displaced the Yellowstone River to the east because the tributaries did not head at the ice front, and those on one side had no advantage over those on the other. A more serious objection is that the asymmetry dates back at least to Flaxville time (Miocene-Pliocene?) as indicated by the far more extensive remnants of the Flaxville terraces on the left side of the present valleys than on the right side.

Homoclinal shifting is an inadequate explanation because the master streams are independent of structure, except locally. The Missouri, for example, cuts across the Poplar dome and the Nesson anticline, and the Yellowstone cuts across the Cedar Creek anticline.

Deflection due to terrestrial rotation seems a possible explanation in the present area but Alden's map of the Number 2 (Cartwright) bench in the area to the south (Alden, 1932, Pl. 1) shows the Cartwright bench to be more extensive on the left side of the river in some places and more extensive on the right in others. That is, Yellowstone River was not asymmetrically disposed in its valley at least in Cartwright time south of the area herein considered.

Progressive regional tilting may be the explanation. Uplift in the northwest would be required to explain the combined displacements of the Missouri and Yellowstone. However, the uplift cannot be related entirely to isostatic adjustment during deglaciation because, as indicated earlier, the asymmetry already existed in Miocene-Pliocene (?) time. Uplift related to the regional tectonics of the western region may be indicated. Localized warping might explain the reversal of asymmetry in Yellowstone Valley south of the map area.

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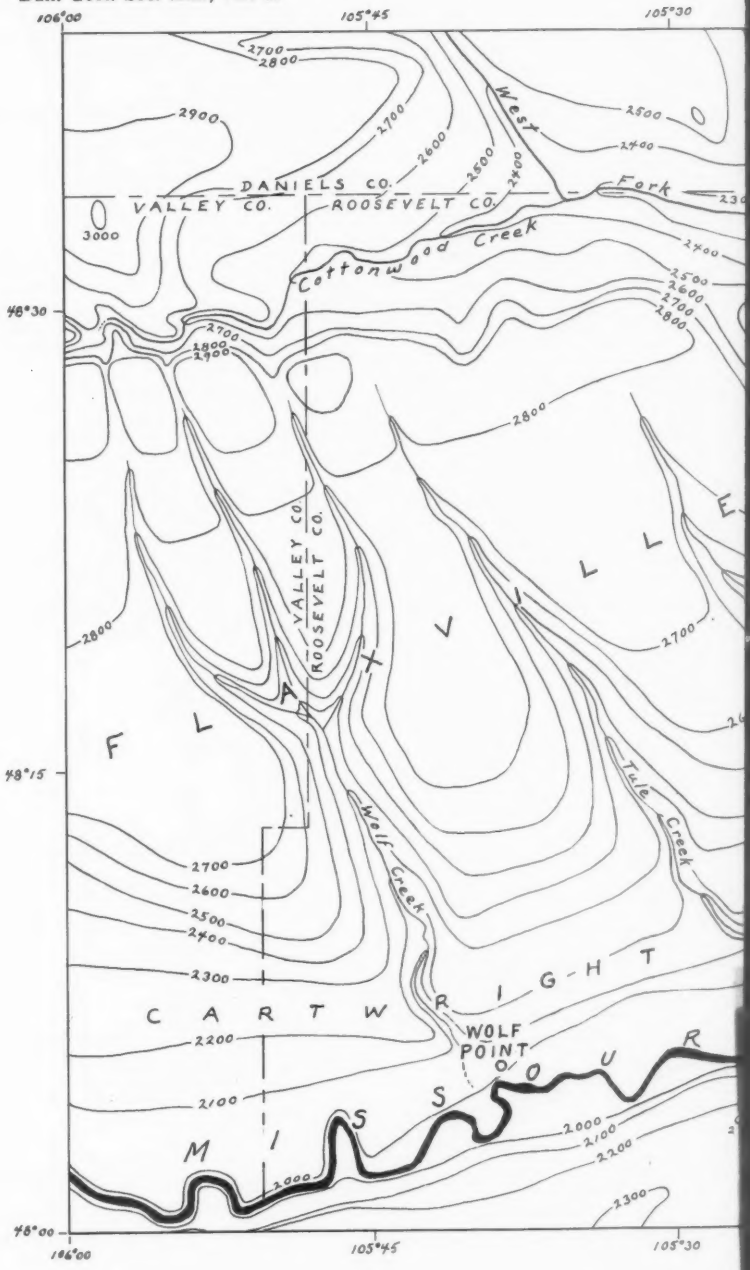

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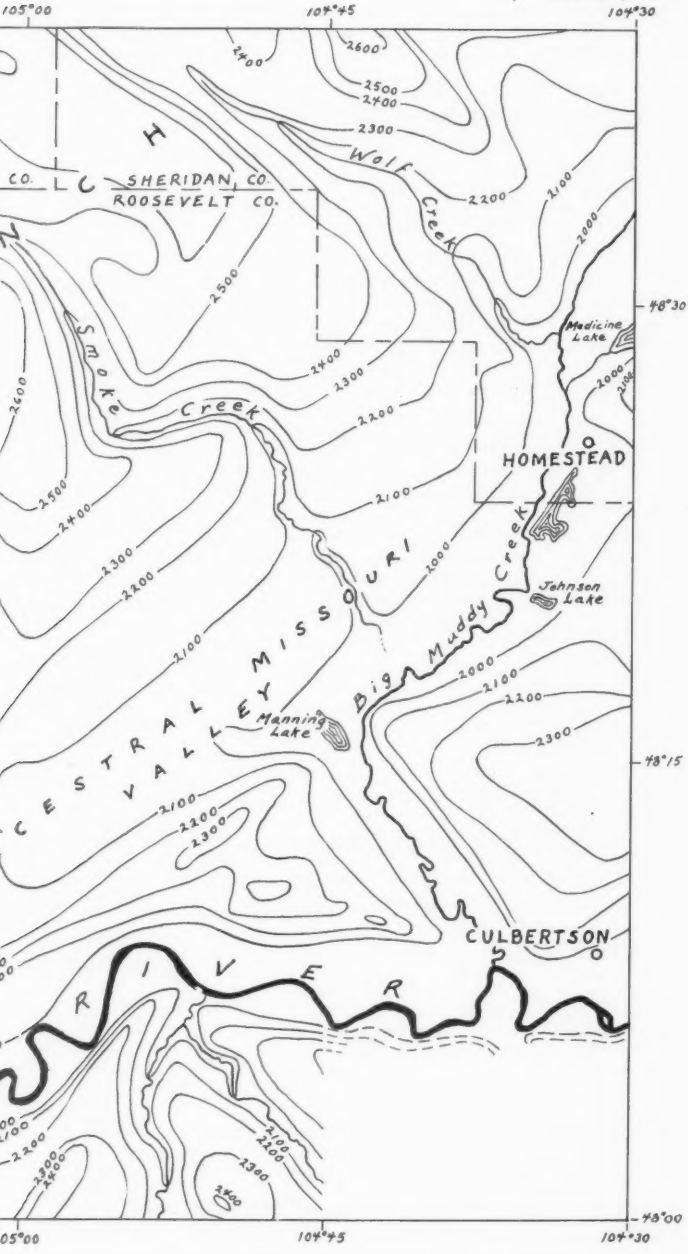

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GLACIAL CHRONOLOGY OF THE NEW ZEALAND PLEISTOCENE

BY MAXWELL GAGE AND R. P. SUGGATE

ABSTRACT

Pleistocene glaciation of New Zealand has been recognized since about 1860, but multiple glaciation has gained general acceptance only recently. The glacial history is directly linked with the marine succession at Ross, Westland, where strongly folded early Pleistocene till, varve silts, and gravels conformably succeed marine Pliocene beds and are overlain unconformably by practically undeformed late Pleistocene moraines. Pleistocene culmination of the Kaikoura orogeny resulted in a clear distinction between early and late Pleistocene glacial deposits and probably explains a mid-Pleistocene hiatus in the successions.

In the proposed chronology, early Pleistocene glacial deposits are referred to the Ross glacial stage. Tectonically undeformed deposits that have accumulated since the present topography has existed are referred to two glacial stages (Waimaungan and Otiran) of the late Pleistocene separated by a full interglacial stage. Local names are given to four distinct advances of substage value recognizable within the Otiran stage. Field recognition depends upon differences in ice extent, erosion and weathering of deposits, and differences of altitude due to progressive lowering of river profiles.

There is as yet little basis for overseas correlation. The Otiran stage may extend farther back into post-Sangamon time than Wisconsin in the strict sense; correlation of Waimaungan with Illinoian is reasonable, though speculative.

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INTRODUCTION

Historical Review

This paper summarizes progress toward establishing a glacial chronology for the Pleistocene in New Zealand. Because it is mountainous and oceanically isolated in the temperate latitudes of the southern hemisphere, New Zealand will probably contribute an important part to the world picture of Pleistocene climatic oscillations.

Geological exploration began in New Zealand about the middle of the nineteenth century, when the glacial theory was still meeting with opposition. The discovery in about 1860 of abundant evidence of glaciation was therefore received with considerable interest. Correlation with the northern hemisphere was hampered by geographic isolation and by the primitive state of New Zealand paleontology, but this did not prevent controversy as to whether the glaciation was Pliocene, Pleistocene, or both. The

extent and number of glaciations also were argued, sometimes bitterly, during the late nineteenth and early twentieth centuries, but on slender evidence and with much confused thinking. With the progress of regional geologic surveys, the distribution and relationships of glacial deposits became better known, and in recent years some moraines have been studied in detail. The onset of Pleistocene cooling in New Zealand is now precisely dated in terms of the local biostratigraphic classification, and correlated approximately with the lower Pleistocene of Europe (Fleming, 1956). Radiocarbon-dating facilities are available, but scarcity of suitable materials has limited the usefulness of this tool in glacial chronology studies.

For the past 40 years it has been accepted that the glaciation of New Zealand was of the mountain-valley and piedmont types, and that no icecap overspread the land at least during the latter half of the Pleistocene.

The development of ideas regarding the number of glaciations was summarized by Willett (1939). As early as 1875 Hutton advocated two glaciations, placing the earlier one in the Pliocene. McKay (1894, p. 43–46) also saw evidence of two glaciations in Westland, but most geologists held to the view of a single great glaciation. Henderson (1917, p. 101; 1931), however, claimed there had been two advances "separated by a considerable time interval" in Nelson province (1931, p. 160), and Morgan (1926, p. 279) predicted correctly that more than two glacial stages would be determined in Westland. Others who saw evidence for more than one ice advance (*e.g.*, Bell and Fraser, 1906, p. 85–86; Cox, 1926, p. 109; Speight, 1938, p. 159) tended to regard them as part of the recessional history of one glaciation, because interglacial deposits had not been found. Multiple glaciation is now generally accepted. Wellman and Willett (1942a, p. 304; 1942b, p. 208) and Gage (1945, p. 157) described glacial deposits in different parts of Westland in terms of separate advances, which Wellman later arranged in sequence and named (1951, p. 30; 1955a, p. 34). These were not claimed to be of equal value, nor to span the entire Pleistocene. Problems in the dating and correlation of glacial deposits in New Zealand were discussed recently by Wellman (1955b) and Gage (1957, mimeographed abstract C22, Australian and New Zealand Assoc. Adv. Sci., 32d meeting).

Climatic History

None of the authors cited above has suggested as many alternations of climate as are recognized in northern hemisphere multiple-glaciation chronologies, yet it is unlikely that the Pleistocene climatic history of the southern hemisphere is less complex than that of the northern. Observations during the past 100 years have shown that mountain glaciers in New Zealand, as elsewhere, reacted to minor climatic fluctuations, and they undoubtedly would have responded even more noticeably to climatic changes of greater magnitude and longer duration such as are believed to have affected the northern hemisphere. Glacial deposits and glaciated landforms must, however, be well preserved if they are to provide evidence for recurrent glaciations and minor oscillations within individual glaciations. Pleistocene diastrophism in New Zealand acted in some ways favorably and in other ways adversely for the preservation of this evidence. The complete glacial history of New Zealand, if known, would probably record more climatic fluctuations than are indicated in the chronology now proposed.

Relationship of Marine and Glacial Successions

Owing to differential tectonic movements and accompanying erosion, Pleistocene sediments provide a somewhat disjointed record of floral and faunal response to temperature fluctuations, but the climatic threshold of the Pleistocene can be defined paleontologically in some unglaciated areas (Fleming, 1953, p. 108; 1956, p. 925–928; Couper and McQueen, 1954). In the Wanganui district of the North Island (Fig. 1), lower Pleistocene fossiliferous marine strata (Nukumaruan and Castlecliffian stages) conformably overlie Pliocene beds and are succeeded, unconformably in the type area, by terrace deposits and littoral beds (Hawera Series), which contain marine fossils in some places. This sequence contains no direct evidence for glaciation, but it provides ecological evidence of cooling from the subtropical temperatures of the Pliocene and of subsequent temperature alternations.

Glacial deposits are widespread in the South Island, but they cannot be linked directly with the nonglacial biostratigraphic classification except at a few Westland localities. The oldest known glacial beds at Ross occur in conformable sequence with Pliocene (Waitotaran) and lower Pleistocene (Nukumaruan) beds (Gage, 1945, p. 143–148; Couper and McQueen, 1954, p. 400–401). This succession is overlain unconformably by younger glacial beds which are middle to upper Pleistocene (Hawera series) on the evidence of plant remains in basal carbo-

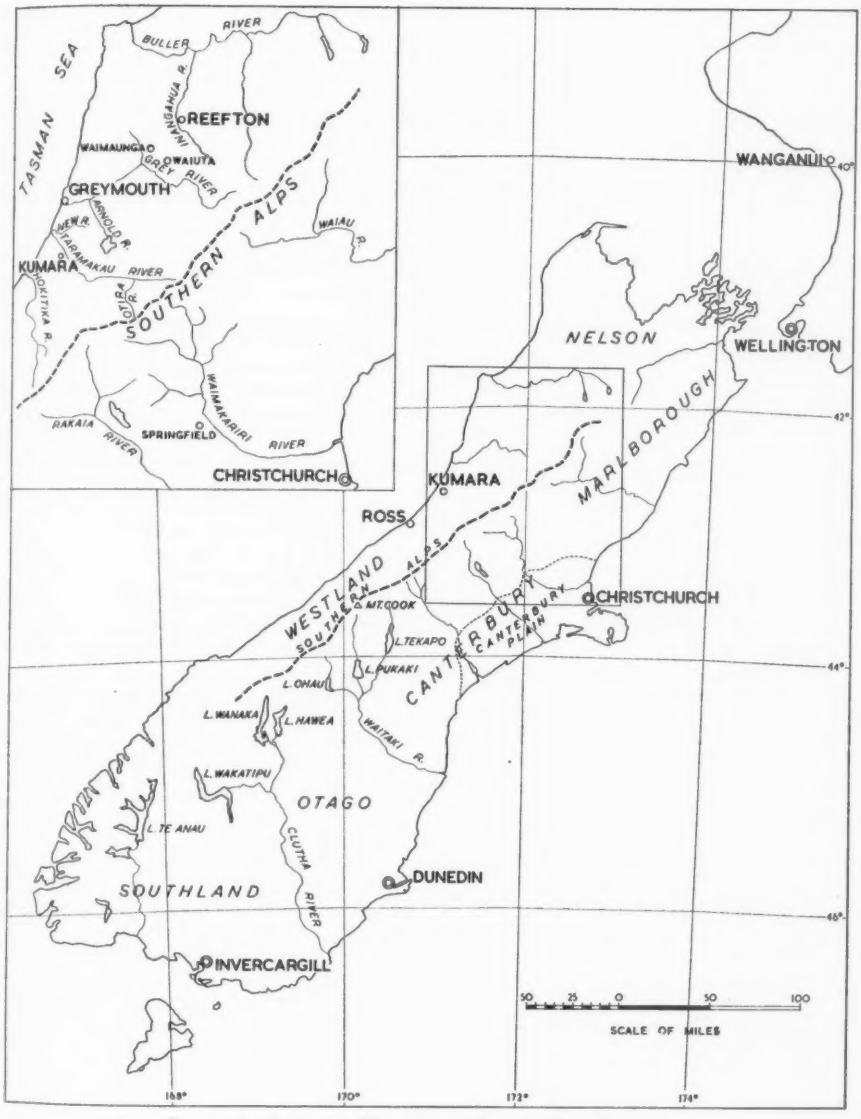

FIGURE 1.—LOCALITY MAP, SOUTH ISLAND, NEW ZEALAND

naceous silts at Ross (Couper and McQueen, 1954, p. 404) and of fossiliferous interstadial marine clays at two other localities (Wellman, 1951, p. 27; 1955a, p. 32–33; Fleming, 1956, p. 927–929).

Tectonic Distinctness of the Lower Pleistocene

Crustal movements that began late in the Tertiary and culminated in the Pleistocene are known in New Zealand as the Kaikoura orogeny (Cotton, 1916, p. 248). The lower Pleistocene glacial beds at Ross were involved in these movements and are therefore distinguishable from later glacial deposits formed since the crisis of the orogeny which show little or no tectonic deformation. Minor local warping continuing into the later Pleistocene contributed to progressive lowering of base levels and in

places has helped to distinguish between the deposits of the later ice advances.

Little evidence survives of the paleogeography at the time of the lower Pleistocene glaciation, whereas the topographic setting of the younger glacial deposits was little different from the landscape of the present day.

Synchronism of Glaciation

The main ice advances were probably distinct and roughly synchronous throughout the glaciated regions of New Zealand. Present-day glaciers of the Southern Alps are nourished by precipitation from predominantly westerly winds, and there is nothing to suggest that snowfall distribution formerly differed from that of today, at least during the latter part of the Pleistocene. Geomorphic evidence indicates that certain present-day rain-shadow areas were formerly unglaciated areas of low precipitation. Willett (1951) estimated that snow line was lowered uniformly by about 3500 feet and that mean temperature was about 6°C. lower, but these estimates do not relate to any specified stage of the Pleistocene. The greatest glacier extensions occurred where the largest glaciers exist now.

As in other countries, glaciers in New Zealand have shrunk considerably since the end of the nineteenth century (Gage, 1951a; 1951b; Harrington, 1952; McKellar, 1955; Suggate, 1950, p. 426–428; 1952, p. 299–301). The response by individual glaciers differs in detail, and the steep rapid-flowing ice streams descending west of the Southern Alps show effects of superimposed short-period fluctuations which are difficult to discern in east-flowing glaciers, but the over-all behavior has been uniform, and climatic fluctuations of a much larger order probably would have resulted in uniform responses throughout the glaciated areas. If this were true, the relative spacing of end moraines marking the maxima of different advances should be similar for all glaciers. This working hypothesis does not conflict with field evidence.

Proposed Glacial Chronology

Type Area

Suitable type sequences of glacial deposits are available both east and west of the Southern Alps. The larger west-flowing glaciers are fed from snow fields on the highest part of the Alps, where the main divide is not far from the west coast (Fig. 1). Comparatively minor advances by these glaciers carried them beyond the present coast line, so that glacial evidence is less well preserved in South Westland than in North Westland and Southwest Nelson, where the divide swings away from the coast. The interrelation of sea-level changes and glaciations may be studied in areas of North Westland where glaciers just failed to reach the coast.

Suggate studied glaciations in North Westland and Southwest Nelson; Gage is most familiar with glacial geology in Canterbury province and especially in the montane valley of the Waimakariri River (Fig. 1), where only the greater ice advances reached beyond the eastern foothills of the Alps. The authors discovered that comparable sequences of events had been derived independently from work on opposite sides of the Alps. Having confirmed this by joint examinations in the field, the writers attempted to apply their chronological classification of deposits to other glaciated valleys, in some cases less well known to them, on the assumption that major advances and retreats were substantially synchronous throughout the South Island. Encouraging results from these preliminary correlations led to a tentative interregional correlation of glacial stages, which was introduced in a joint contribution to Section C of the Australian and New Zealand Association for the Advancement of Science at the Dunedin meeting, January 1957 (Gage and Suggate, 1957, mimeographed abstract C24, Australian and New Zealand Assoc. Adv. Sci., 32d meeting).

Terminology

Separate advances will require local names in each glacier system as long as doubt remains as to the correctness of correlations, and it may be desirable to retain local names permanently. Nevertheless, a terminology of general application, independent of chronologies based on nonglacial deposits, is needed.

Wellman (1951, Fig. 2) nominally introduced local names for glacial "stages", but discussed them as "advances" (1951, p. 30–32) in an account of South Westland glaciation. For reasons already given, the sequence of glacial deposits in this region is less complete, and the relationships between successive advances are less clear than farther north. The authors therefore consider it advisable to base the new terminology on localities in North Westland and Southwest Nelson.

The following proposed classification of New Zealand glacial deposits is independent of the nonglacial stratigraphic classification, with

which younger glacial deposits are not readily correlated:

Late Pleistocene	OTIRAN GLACIAL STAGE: four principal advances
	Interglacial interval
	WAIMAUNGAN GLACIAL STAGE: not subdivided
	Long interval for which a full record is nowhere available
Early Pleistocene	ROSS GLACIAL STAGE: not subdivided

Basis for Differentiating Stages

In the absence of evidence for lengthy intervals between them, the four most recent major advances are grouped in a single glacial stage, named from the Otira River (Fig. 1), because ice from the Otira headwaters flowed down two valleys (Taramakau and Waimakariri), respectively northwest and southeast of the main divide, in which the authors have studied glacial histories.

A long ice-free interval, as yet unnamed, is indicated by extensive erosion and deep oxidation of pre-Otiran glacial deposits. Pre-Otiran deposits are referred to the Waimaungan glacial stage if they are: (a) tectonically undeformed apart from slight warping and minor faults with physiographic expression; and (b) formed within the broad framework of the present topography. The Waimaungan glaciation is the earliest for which geomorphic evidence survives. The stage is named from Waimaunga, in the valley of the Grey River, Southwest Nelson (Fig. 1). Outwash gravels may be traced back from here as a prominent high-level constructional surface to moraine near the abandoned mining town Waiuta.

Deposits of the Ross glacial stage have been deformed during the Kaikoura orogeny and clearly belong with lower Pleistocene and upper Tertiary rocks.

Ross Glacial Stage

The term Ross glaciation was introduced by Fleming (1956, p. 926) following Wellman (1951, Fig. 2) and was based originally upon glacial beds more than 90 feet thick at Ross, which Gage (1945, p. 146–147) described under the heading "R.7.: Laminated silts and boulder clay." R.7. was considered to be a glacial deposit because it not only resembles till and varve silts formed during later glaciations for which morphologic evidence survives, but also includes angular blocks of rocks for which no source exists within 9 miles. Striated pebbles have since been observed in R.7. The upper part of the underlying formation ("R.6.: Lower decomposed conglomerates", Gage, 1945, p. 144–145) consists of 300–400 feet of coarse gravel including, near the top, large angular blocks of a rock type foreign to the local terrain. Gage now considers these gravels to be proglacial outwash of the advancing Ross ice.

Distinctively glacial characters are absent from the conformably overlying "R.8.: Upper decomposed conglomerate", which is correlated with the upper part of the Old Man Gravels, a formation which attains thicknesses of up to 5000 feet elsewhere in the region. Gage's tentative correlation of varve silts and till at certain other North Westland localities with R.7. beds at Ross (1945, p. 147) is supported by later work (Bowen, 1957, mimeo. abst. C26, Australian and New Zealand Assoc. Adv. Sci., 32d mtg.). Carbonaceous sediments at the base of R.6. yielded a lower Pleistocene (lower Nukumaruan) microflora (Couper and McQueen, 1954, p. 416), marking the transition from marine upper Pliocene (Waitotaran) to the apparently nonmarine Ross glacial beds. The whole succession, including Old Man Gravels, is overlain unconformably at Ross by Otiran glacial beds.

Interval Between Ross and Waimaungan Stages

Old Man Gravels in Westland represent the thick, coarse clastic facies, known by various local names, which in many parts of New Zealand terminates the Tertiary marine succession. Together with the Tertiary rocks, the gravels have been folded and faulted during the Kaikoura movements responsible for the present major tectonic relief. Many references to these coarse beds are scattered through the New Zealand literature, and it is generally held that they are formed of the waste shed from rapidly rising mountain axes during the orogenic crescendo.

Unconformity separates Waimaungan glacial beds from Old Man Gravels in Southwest Nelson (Suggate, 1957, p. 62). The "Piedmont glaciation" of this area forms part of the basis for the Waimaungan stage, and although tilted and uplifted locally, Waimaungan beds were formed after the climax of the orogeny. Circumstances were unfavorable for the survival of mid-Pleistocene glacial evidence, but Couper and McQueen (1954, p. 412, 420) and Fleming (1956, p. 928) record faunal and floral evidence

for post-Ross renewed cooling in upper Nukumaruan sediments from the North Island.

Waimaungan Glacial Stage

Glaciers were thicker and more extensive both east and west of the Alps during the Waimaungan glacial stage than at any later time. Coalescing glaciers in Southwest Nelson advanced northwest into a broad tectonic depression occupied now by parts of the Grey and Inangahua rivers, and the ice almost certainly reached the sea south of Greymouth. Widely distributed remnants of moraine and glaciofluvial gravels south and southeast of Reefton include the end moraine at Waiuta mentioned earlier and outwash extending westwards. Remnants of the constructional surface survive on crests of ridges up to 700 feet above the present river beds. Waimaungan deposits in the Reefton district, mapped by Suggate (1957) as "Piedmont glaciation deposits", are correlated with high-level gravels traceable northwards down the Inangahua Valley to beyond the junction of Inangahua and Buller rivers. Waimaungan gravels and local moraine veneer a high-level valley floor 500 feet or more above the present bed of the Buller River. Suggate now considers that high-level gravels near Ross which Gage failed to recognize as glacial ("R.9."; 1945, p. 145) probably represent the Waimaungan glacial stage in that area.

Waimaungan deposits east of the Southern Alps are best known in the valley of the Waimakariri River (Figs. 1 and 3), where deeply eroded remnants of formerly thick and extensive glaciofluvial beds, till, and varve silts occur at several places. The most extensive remnant underlies a plateau east of Avoca (Fig. 3), beneath which glacial beds appear to extend continuously to river level, nearly 1500 feet below. Avoca was chosen as the local name for the Waimaungan deposits and for the ice advance which produced them (Gage, 1958, p. 129–134). Avoca deposits occupy an overdeepened trough from which a high-level outlet leads out over a sill between the Puketeraki and Torlesse ranges (Fig. 3). The Waimakariri River is now entrenched more than 1000 feet below the floor of this former ice-discharge channel, which during the maximum advance discharged an ice-tongue that expanded eastwards over part of the area now occupied by the Canterbury Plain (Fig. 1.).

No evidence has been found for major oscillations within the Waimaungan glacial stage.

Waimaungan-Otiran Interglacial Interval

Downcutting by rivers and deep oxidation of Waimaungan deposits preceded the next glacial advance. Avoca beds in the Waimakariri Valley are oxidized at least superficially throughout, except for the relatively impermeable varve silt and till layers at the base. Weathering of west coast Waimaungan deposits is even more noticeable, perhaps partly because of regional climatic difference, but also because of a wider range of lithology, including a higher proportion of less-resistant rock types than are present east of the main divide. The status of a full interglacial stage is justified for the Waimaungan-Otiran interval on grounds of the amount of river downcutting (1000 feet in the Waimakariri) and the great depth of oxidation, which seems more than would be expected from the cumulative effect of intra-Otiran interglacial breaks and postglacial weathering.

It must be emphasized that, in contrast with continental glacier deposits, there is here no simple and direct relationship between depth of weathering profile and the age of the deposit, although the intensity of weathering of the whole deposit appears to be related in a general way to its age. Thus, permeable Ross deposits are thoroughly weathered, Waimaungan and the earliest Otiran are yellowish and moderately oxidized, and later Otiran are fresh except for a shallow oxidized zone, the depth of which varies with slope and other factors. The great depth of oxidation of the older deposits may be explained by great vertical fluctuations of water table, resulting from changes of base level, or of climate, or both. The present water table beneath the piedmont alluvial apron of the Canterbury Plain is more than 300 feet below the surface in places, in a region of temperate, humid climate; elsewhere it is very near the surface.

Otiran Glacial Stage

Local names for the ice advances grouped as the Otiran glacial stage are shown with suggested correlations. Certainty of correlation decreases with increasing age.

Westland and Southwest Nelson		*Waimakariri Valley (Canterbury)*
Kumara-3	Otiran glacial stage	Poulter
Kumara-2		Blackwater
Kumara-1		Otarama
Hohonu		Woodstock

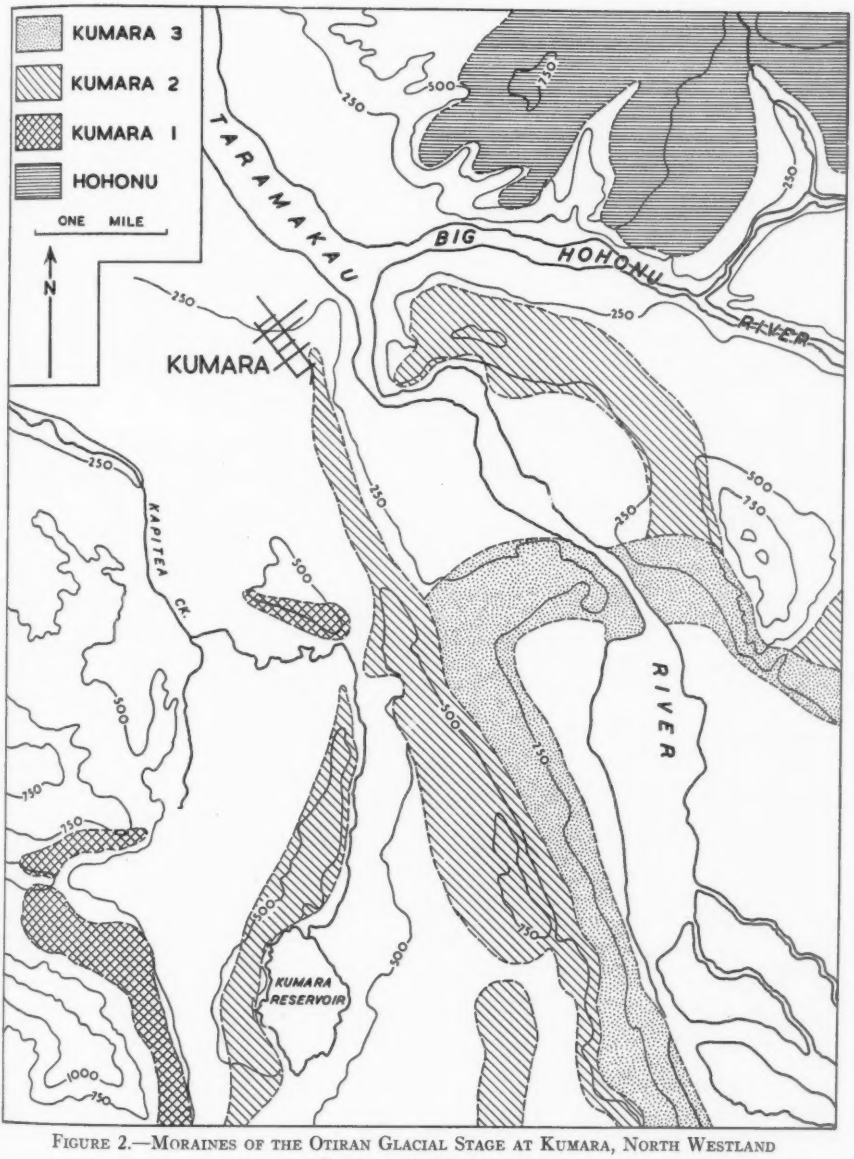

FIGURE 2.—MORAINES OF THE OTIRAN GLACIAL STAGE AT KUMARA, NORTH WESTLAND
Contour values in feet.

In both areas, the deposits from the first Otiran advance are more weathered than those from the last three, which form an obvious subgroup in each case. Otiran deposits in several localities rest unconformably on river- or ice-eroded surfaces cut in Waimaungan deposits.

The limits of the Hohonu advance are clearly defined by lateral and end moraines north of Big Hohonu River (Fig. 2). The Hohonu end moraine northeast of Kumara is completely separated from lower-level moraines and outwash produced by less extensive Kumara advances, but its general setting and fresh

glacial topography place it in the Otiran rather than in any earlier stage. The first two Kumara advances were similar in extent, but their deposits can be distinguished by differences of elevation and by the deeper weathering of Kumara-1 outwash. Marine erosion has cliffed Kumara-1 and Kumara-2 outwash at different levels a few miles farther west. Kumara-3 glaciers were distinctly less extensive than those of Kumara-2 and Kumara-1. Kumara-2 ice in the Grey valley (Fig. 1) blocked a tributary and dammed a lake which was eventually filled in by outwash from other tributaries. Sediments from this lake contained plant stems giving a radiocarbon age of 22,300 ± 350 years. (Ref. 14CNZ/S44/828; coll. E.T. Annear).

The Woodstock advance in the Waimakariri Valley (Fig. 3) reached the mountain front and is marked by moraine and outwash that are appreciably more weathered than those of later substages (Gage, 1958, p. 136–138). Outwash gravel from the Waimakariri and adjoining glaciated valleys coalesced to form a piedmont alluvial plain comparable with the present Canterbury Plain but at a higher level at its inland margin close to the foothills of the Southern Alps. This former plain was deeply and widely eroded before the succeeding Otarama advance. The western portion of the Waimakariri segment of the Canterbury Plain is a constructional surface upon Otarama outwash. Although Otarama ice temporarily thrust forward as far as the inner margin of the Canterbury Plain, this substage was perhaps not much more important, in terms of ice volume, than the succeeding double Blackwater advance, which failed to reach as far as the mountain front. The final (Poulter) advance was less extensive and produced far less outwash than preceding advances. Profiles constructed on the basis of remnants of moraine and outwash from the several Otiran advances enable glaciations of the upper Waimakariri Valley to be linked with the history of the Canterbury Plain (Gage, 1958, Fig. 3).

Correlatives of the later Otiran substages have been identified tentatively in South Westland. The Paringa and Kinnaird advances (Wellman, 1951) are referred to Kumara-1 and Kumara-2, and the Alpine advance to Kumara-3. Similarly on the east side of the Southern Alps, the innermost moraine loops at the southern ends of lakes Tekapo, Pukaki, Ohau, Hawea, and Wanaka (Canterbury and Otago; Fig. 1) are considered equivalent to the Poulter substage, and earlier Otiran deposits also are recognized.

Minor Post-Otiran Readvances

Small end moraines at present river-level survive in some valleys closer to existing glaciers than to late-Otiran end moraines. These are ascribed to a short-lived late advance, perhaps between 500 and 2000 years ago. Small vegetation-covered moraine loops near the late nineteenth century maximum of the larger existing glaciers indicate a more recent advance, from which there has been irregular shrinkage for more than 200 years (Harrington, 1952). A small moraine loop near the snout of the Franz Josef glacier is associated with an unvegetated trim line, which in the Westland rain forest environment is a certain indication of youth.

Status and Overseas Correlation of New Zealand Stages

Radiocarbon dating has so far provided little support for the chronology discussed in this paper, chiefly because of the rarity of carbonaceous material in deposits whose position in the sequence of stages and substages is clear. Moreover there is little prospect of direct paleontologic correlation of New Zealand glacial sequences with those of the northern hemisphere. Nevertheless, applying the kind of tests used to differentiate glacial stages in other parts of the world, it seems certain that the Waimaungan and Otiran deserve the status of separate stages. Reasons have already been given for considering the Waimaungan-Otiran interval to be of stage magnitude.

The question of whether to attempt identification of these stages with those defined in the northern hemisphere remains. The Otiran stage is clearly made up of a single important early substage and a group of three later substages. The degree of weathering of the first substage (Woodstock; Hohonu) and the topographic changes that intervened before the next substage (Otarama; Kumara-1) suggest that the earlier one is perhaps of the order of twice as old. It is thus probably older than any of the original substages of the Wisconsin, which are spaced between 25,000 years and 8000 years ago (Horberg, 1955), but may be compared with the deposits of post-Sangamon, pre-"type"-Wisconsin age discussed by Flint (1956). The single uncomplicated radiocarbon date obtained so far (Kumara-2: 22,300 ± 350 years) supports this comparison.

Although there can be little doubt that the Wisconsin and at least the latter half of the

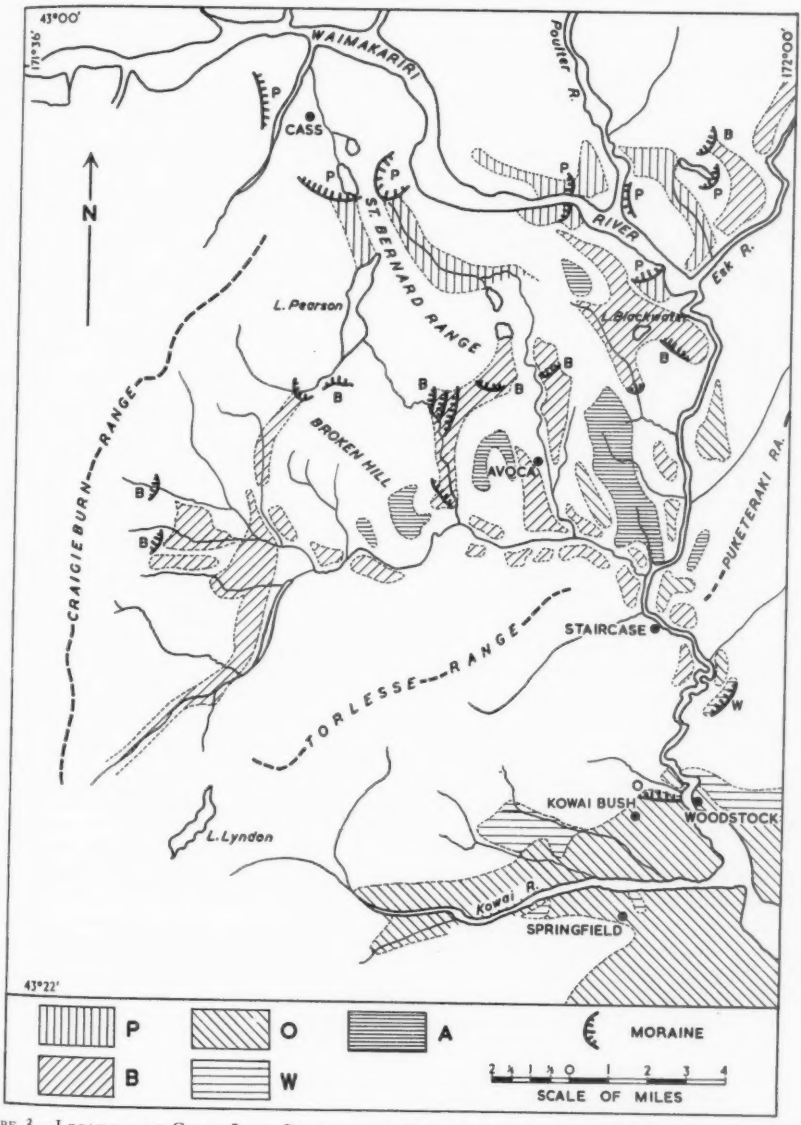

FIGURE 3.—LOCATION OF CHIEF LATE PLEISTOCENE GLACIAL DEPOSITS IN THE WAIMAKARIRI VALLEY, CANTERBURY

Waimaungan and Otiran deposits (till, ice-contact sediments, moraine, outwash) differentiated according to age by rulings. In order of decreasing age: *A* = Avoca glaciation (Waimaungan glacial stage); *W*, *O*, *B*, *P* = Woodstock, Otarama, Blackwater, Poulter advances (Otiran glacial stage). Moraines shown include chief end and recessional moraines and some prominent lateral moraines.

Otiran span the same interval of time, the terms are not intended to be equivalent, and correlation of Otiran substages with those of the Wisconsin is not justified at present. Correlation of Waimaungan with Illinoian is reasonable though speculative, but grounds are far less secure for identifying the Ross glacial stage with Nebraskan or Günz.

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Short Notes

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Sr/Rb AGE MEASUREMENTS ON HORNBLENDE AND FELDSPAR, AND THE AGE OF SYENITE AT CHICOUTIMI, QUEBEC, CANADA

BY W. H. PINSON, JR., H. W. FAIRBAIRN, AND R. F. CORMIER

Age measurements on coexisting minerals in igneous rocks are of particular interest in geologic time studies. We are reporting here, as far as we are aware, the first such measurements that include hornblende. The preferred minerals for Sr/Rb mass spectrometric analysis are biotite and muscovite. These micas, which have favorably small Sr/Rb ratios, are in many cases absent (or, in the case of biotite, chloritized) from Precambrian rocks. Hornblende may in some cases be an invaluable substitute, as our example indicates.

Table 1 shows analytical data for the hornblende and feldspar extracted from syenite from Chicoutimi, Quebec. The notation H3070 refers to hornblende extracted from sample No. 3070, and F3070 to feldspar from 3070, and the rock samples are described in a footnote to Table 1. Two additional analyses of hornblende from younger rocks are included as examples with unfavorably large Sr/Rb ratios for age determination. To avoid unsuccessful mass spectrometric work it is advisable first to carry out an exploratory spectrochemical analysis on the hornblende samples. This precaution is usually unnecessary for the potassium feldspar from acidic rocks, if a carefully adjusted heavy liquid is used to float the potassium-rich fraction of the feldspar. For example, although F3070 contains only 9.4 per cent potassium (flame photometric analysis), compared to a possible 14 per cent for the pure compound, the untreated feldspar concentrate contains even less potassium and is too calcic (Sr-rich) for Sr/Rb analysis.

Separate Sr isotope-ratio analyses were made for the hornblende H3070 and the feldspar F3070, and ages have been calculated from both the isotope-dilution (I.D.) and the isotope-ratio (I.R.) data. Both normal and radiogenic Sr can be simultaneously measured by a single isotope-dilution analysis, but, unless there is a large excess of radiogenic Sr^{87} over normal Sr^{87}, there will be a relatively large associated error. Thus, in the cases of potassium feldspars and hornblendes in which the ratio of normal Sr^{87} to radiogenic Sr^{87} is large, separate isotopic-composition analyses of the extracted, unspiked Sr in addition to the isotope-dilution analysis greatly increase both precision and accuracy of the age determination. The rather small and favorable Sr/Rb ratio of H3070 indicated that a separate isotope-ratio analysis was not necessary, but the analysis was made to demonstrate this point. The unfavorably larger Sr/Rb ratio in F3070 indicated the need for a separate isotope-ratio analysis, and this analysis greatly reduced the error in age.

Details of the isotope-ratio analyses are given in Table 2, and the composition of "normal" Sr (Nier, 1938) is given for comparison. The measured ages are reported to the nearest 10 million years. The variations in the stable-isotope ratio 86/88 from "normal" may or may not be real, and these small differences of 1 per cent or less in the 86/88 ratio from normal should not be taken as meaningful. Some of the variation probably arises from filament fractionation and other instrumental and measurement errors. The greatest error is expected in the 84/88 ratio, because of the 4½ per cent difference in mass. The 87/86 ratio is more trustworthy than either the 84/88 or 86/88 ratios, because the mass difference is smaller, and the measurement is made on the same sensitivity scale (100 millivolt scale of a vibrating reed electrometer). The 84/88 ratio in F3070 is normal, and this ratio is not reported for H3070 because the 84 peaks, taken on the 10 millivolt scale, were too short for accurate measurement.

The results shown in Table 1 warrant further Sr/Rb age studies of coexisting hornblende and feldspar in mica-poor rocks. Aldrich (1956), working on the Brown Derby pegmatite and

the Quartz Creek granite near Gunnison, Colorado, has demonstrated the usefulness of the Sr/Rb method applied to potassium feldspar. Since, in the present pair of analyses, the feldspar age is the lower, it would be of interest to have further data elsewhere. Aldrich (1955) has already noted that Sr/Rb feldspar ages are somewhat of the order of 5 per cent lower than coexisting mica ages, though the difference is not uniform.

The age of 900 × 10^6 for the Chicoutimi syenite agrees closely with an age measurement by Nier (1939) on cleveite from Pied des Monts, Quebec, a locality about 150 miles east-northeast of Chicoutimi, in which a 206/238 age of 882 × 10^6 years and a 207/206 age of 905 × 10^6 years were reported. This is the nearest locality on which an age measurement has been made with which the Chicoutimi syenite may be compared. The Chicoutimi syenite age is also within the spread of ages from scattered localities usually considered to be in the Grenville age province. For example, Tilton (1955) reported 1050 × 10^6 years on zircon and sphene from granite between Tory Hill and Essonville in the Haliburton area of Ontario. Other ages from the Grenville age province are summarized by Faul (1954) and are in the neighborhood of 1000 × 10^6 years. Thus, the syenite from near Chicoutimi, Quebec, on the basis of these age measurements and the stated limits of error, can possibly be assigned to the same orogeny which produced the Grenville rock metamorphism.

TABLE 1.—ANALYTICAL RESULTS AND AGES

No.	Rb	Sr (Normal)	$*Sr^{87}$ I.D.	$*Sr^{87}$ I.R.	Age I.D.	Age I.R.
					(10^6-year units)	
H3070	72	10.3	0.29	0.27	1020 ± 100	950 ± 50
F3070	157	67.3	0.37	0.52	600 ± 300	840 ± 40
H3069	106	1220	−0.03	.....	.....	.....
H3136	13.5	26.7	0.00	.....	.....	.....

3070, Syenite from Route 54, 300 yards south of junction with 16A, near Chicoutimi, Quebec, collected by H. W. Fairbairn.

3069, Tinguaite from St. Joseph's Blvd., Montreal East, collected by H. W. Fairbairn.

3136, Granite from Oak Bay, New Brunswick, collected by J. B. Thompson, Jr.

Analyses in ppm by weight. λ Rb = 1.39 × 10^{-11} $year^{-1}$ (Aldrich *et al.*, 1956). Assumed initial relative abundance of Sr^{87} is 0.0696. Weight fraction of Rb^{87} is 0.2832. Age = $(1/\lambda Rb)$ ln $(1 + [*Sr^{87}/Rb^{87}])$.

TABLE 2.—Sr ISOTOPE RATIO MEASUREMENTS

No.	87/86	84/88	86/88
H3070	0.977 ± 0.003	.....	0.1180 ± 0.0012
F3070	0.792 ± 0.002	0.0068 ± 0.0002	0.1185 ± 0.0005
Normal Sr (Nier, 1938)	0.712	0.0068	0.1194

Acknowledgments

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AGE OF THE METAMORPHISM OF THE ROCKS OF THE MANHATTAN PRONG

BY LEON E. LONG AND J. LAURENCE KULP

The igneous-metamorphic complex in the vicinity of the city of New York consists of two major subprovinces: the New Jersey-New York Highlands, an elongated belt trending northeast from western New Jersey to northwestern Connecticut (Fig. 1), and the Manhattan Prong lying south of the Highlands. The rocks of the Prong have been classified by some workers into the Fordham gneiss, overlain in turn by the Lowerre quartzite, Inwood marble, and Manhattan schist and gneiss. This is a lithologic sequence strikingly analogous to the rocks north of the Highlands, where Precambrian gneisses are overlain by a fossiliferous Cambro-Ordovician series of Poughquag quartzite-Wappinger dolomite-Hudson River pelite.

For many years there have been discussion and debate as to whether the highly metamorphosed rocks constituting the Prong are Paleozoic or Precambrian. Berkey (1907) believed that, since the Poughquag quartzite lies unconformably upon the gneisses north of the Highlands, and the Inwood is essentially conformable with the Fordham in the south, there is no compelling reason to correlate these two sequences across the Highlands barrier. He therefore favored the opinion that all the Prong rocks are Precambrian. On the other hand Balk (1936), on the basis of his classic study in Dutchess County, New York, concluded that the Cambro-Ordovician sequence could be traced eastward through rapidly increasing grades of metamorphic intensity into a sequence identical to that of the Prong around New York, and that there is no fundamental break between the Dutchess County rocks and the highly metamorphosed rocks farther south. He found evidence that any unconformity which may have existed between the Fordham and the Inwood (the intervening Lowerre quartzite is present in only a few localities) was converted into a pseudo-conformity by the intense deformation and shearing. More recently Paige (1956) correlated undoubted Cambro-Ordovician rocks west of the Hudson River near Peekskill, New York, with Inwood marble and Manhattan schist east of the river.

The potassium-argon method of age determination makes it possible to measure the time of the latest metamorphic event. Although such measurements cannot define the age of the original sedimentation, they can help clarify the geologic history of the area. Potassium-argon ages on biotite, phlogopite, and muscovite from several different rock units and locations in the Manhattan Prong are reported here (Table 1; Fig. 1). The experimental techniques were those of Carr and Kulp (1957) as modified by Damon and Kulp (1957). More than a dozen replicate analyses of micas in the 300–500 m.y. age range have established the experimental reproducibility as approximately 4–5 per cent. The radiogenic argon concentration in all samples reported here was greater than 92 per cent.

The measured ages found for the metamorphism of the schist and marble and for the pegmatites are in excellent agreement. The standard deviation for seven out of eight samples lies within the uncertainty of the experimental measurements. The agreement among these diverse mineral types suggests that the average apparent age, 366 ± 9 m.y., is close to the true age for the latest major metamorphic event. It should be noted that similar ages ranging from 320 to 380 m.y. have been reported for pegmatites, granites, and metasediments all along the Atlantic seaboard (Wasserburg *et al.*, 1957; Eckelmann and Kulp, 1957; Carr and Kulp, 1957; Kulp, Long and Eckelmann, 1957). This age is tentatively correlated with the upper Ordovician, Taconic orogeny.

The two samples of biotite from the Fordham gneiss appear to be significantly older. This may be interpreted in two ways: (1) the Fordham represents older basement rock, on which the Lowerre, Inwood, and Manhattan formations were deposited, which did not lose all of its argon during the 365 m.y. recrystallization, or (2) the biotite contained some impurity such as amphibole which may carry excess argon. Preliminary results on rubidium-strontium age determinations on Manhattan Prong rocks indicate an age of 350–400 m.y., including the Fordham gneiss, but the small ratio of rubidium/common strontium in the particular samples investigated made the error sufficiently large to make it impossible to eliminate one of these

hypotheses. The phenomenon of argon loss from older rocks during superimposed metamorphic events has been clearly demonstrated in the Lamont laboratory for the Spruce Pine, North Carolina, area and is suggested by preliminary work along the contact of the Manhattan Prong with the New Jersey-New York Highlands.

It is concluded that the age work thus far done suggests that the latest major orogenic event that produced the metamorphism of the rocks of the Manhattan Prong and the intrusion of the pegmatites occurred about 365 m.y. ago and that this result is not inconsistent with a Cambro-Ordovician date for the deposition of the Lowerre, Inwood, and Manhattan formations.

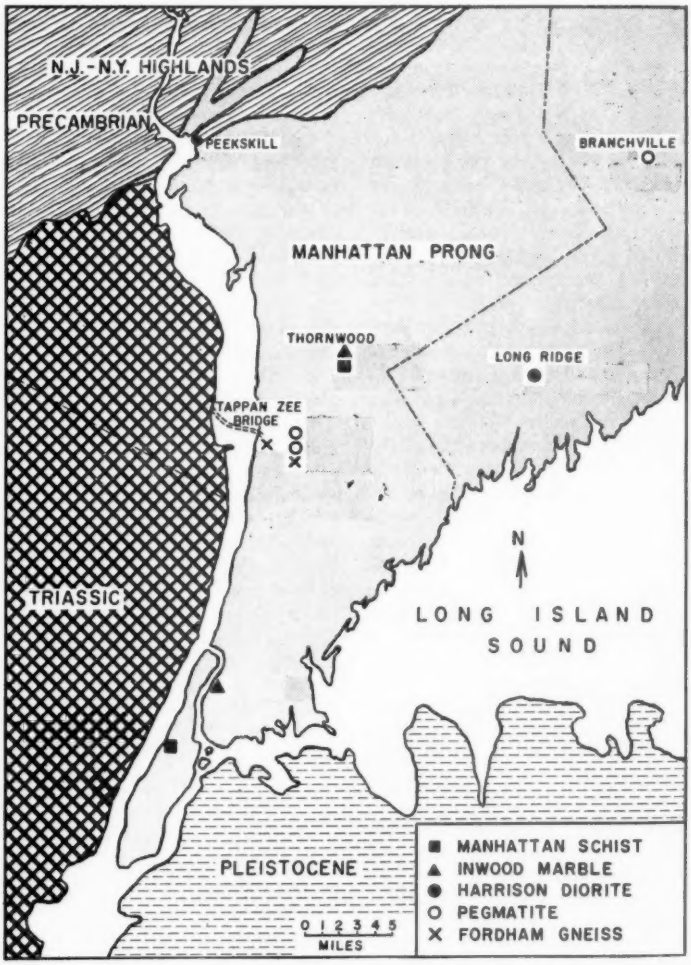

FIGURE 1.—LOCATION OF SAMPLES FOR POTASSIUM-ARGON AGES, MANHATTAN PRONG

Further work in the city of New York area is in progress using potassium-argon, rubidium-strontium, and uranium-lead methods and will be reported later.

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TABLE 1.—MANHATTAN PRONG POTASSIUM-ARGON AGES*
(Probable uncertainty = 4–5% of the apparent age)

Sample location	Mica type	K(%)	Radiogenic A^{40} (ppm)	A^{40}/K^{40}	Apparent age (m. y.)
Inwood marble, River Ave., Bronx, N. Y.	Phlogopite	8.14	0.206	0.0208	(325)†
Manhattan schist, Riverside Church, New York, N. Y.	Biotite	5.00	0.141	0.0232	359
Manhattan schist, railroad station, Thornwood, N. Y.	Biotite	3.82	0.114	0.0244	376
Inwood marble, quarry, Thornwood, N. Y.	Phlogopite	7.09	0.194	0.0225	349
Harrison diorite, Long Ridge Road, Rte. 104, ∽ 3 miles south of Long Ridge, Conn.	Biotite	7.69**	0.228	0.0243	375
Branchville pegmatite, quarry, Branchville, Conn.	Muscovite	8.49**	0.243	0.0235	364
Discordant pegmatite intruded into Fordham gneiss, 2.3 miles east of Tappan Zee Bridge, N. Y. Thruway	Biotite	7.76	0.226	0.0239	368
Discordant pegmatite intruded into Fordham gneiss, location above	Muscovite	8.47	0.248	0.0240 Average	371 366 ± 9 s.d.
Fordham gneiss country rock, location above	Biotite	6.69	0.213	0.0262	400
Fordham gneiss country rock, 0.6 mile east of Tappan Zee Bridge, N. Y. Thruway	Biotite	5.78	0.210 0.199	0.0299 0.0282	440

* Calculated using $\lambda_\beta = 4.8 \times 10^{-10}$ yr.$^{-1}$ and $\lambda_e = 0.58 \times 10^{-10}$ yr.$^{-1}$.

† Sample badly weathered; age not included in the average.

** Analyzed by D. Thaemlitz and C. O. Ingamells, Rock Analysis Laboratory, Univ. of Minnesota; others by Ledoux and Co., Teaneck, N. J.

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SUTURE OF *ACANTHOHOPLITES ASCHILTAENSIS* (ANTHULA)

BY ALEXANDER STOYANOW

Acanthohoplites Sinzow (1907, p. 456–458), type genus of Acanthohoplitinae, was introduced for those forms grouped under *Parahoplites aschiltaensis* Anthula (1899, p. 110), which are characterized by the ribs bearing tubercles, essentially at the point of bifurcation, and by the suture with a symmetrical first lateral lobe, as distinct from *Parahoplites, sensu stricto*, represented by *P. melchioris* Anthula (1899, p. 112, Pl. 8, figs. 4a–5b), a genus with untuberculated ribs and an asymmetrical first lateral lobe.

Acanthohoplites aschiltaensis (Anthula, 1899, p. 117, Pl. 10, figs. 2a–2b, 3a–3b, 4; Pl. 11, fig. 1), genotype by a subsequent designation (Stoyanow, 1949, p. 106), originally was illustrated in four specimens of which drawings only two agree with Anthula's diagnosis in the text: one, of 70 mm diameter, Figures 2a–2b of Anthula's Plate 10, and the other, of 45 mm diameter, Figures 3a–3b of Plate 10. The latter example was designated as the holotype of Anthula's species for two reasons: (1) as illustrated, it clearly shows the tuberculation of predominantly bifurcating primary ribs characteristic for the genus; (2) it is the only specimen of Anthula's types with a figured suture.

As has been pointed out (Stoyanow, 1949, p. 98, 106–109), this suture, Figure 3b of Anthula's plate 10, re-illustrated by Roman (1938, p. 350, Fig. 331 only) and by Moore *et al.* (1957, p. L386, Fig. 4c), and reproduced again in Figure 3 of Plate 1 of this paper, does not conform with the suture of numerous species referable to *Acanthohoplites* from the Lower Cretaceous strata of the Caucasus, Mangyshlak, and Arizona (Sinzow, 1907, Pl. 4, figs. 11, 15, 16, 21; Pl. 6, figs. 1, 2; Pl. 7, figs. 1, 2; Pl. 8, fig. 1; Stoyanow, 1949, Pl. 19, figs. 8, 23; Pl. 20, figs. 6, 16; Pl. 25, fig. 4; *see also* Figures 6 and 7 of Plate 1 of this paper). As discussed, the essential difference is in the asymmetrically bifid first lateral lobe in the suture of the genotype, as illustrated, which so markedly disagrees with the strongly symmetrical tripartite first lateral lobe of the suture in the described congeneric species. This discrepancy was attributed either to the imperfect preservation of Anthula's type or to the incorrectness of the original drawing. However, his frequent allusion to the similarity between the suture of his species, since interpreted as *Acanthohoplites*, and that of *Parahoplites*, was also mentioned. This similarity was strongly opposed by Sinzow (1907, p. 479) in his fundamental formulation of the difference between the suture of these two genera:

> "In der Fig. 3, Taf. 5 ist das Fragment einer feinrippigen Varietät [of *Acanthohoplites aschiltaensis*] abgebildet, ebenso in Fig. 2 dieser Tafel. Auf dem ersten sind die Suturlinien gut erhalten, die im Gegentheil zur Behauptung Anthula's (loc. cit., S. 117) gar nicht denem von *Parahoplites Melchioris* änlich sind, da die oberen Lateralloben bei letzeren gar keine Asymmetrie zeigen, die so charakteristisch für die gattung *Parahoplites* ist."

Obviously, the word "letzeren" in the last sentence pertains to *Acanthohoplites* and not to *Parahoplites*. We shall return to this interesting fragment later. It was manifest that the nature of the suture in *Acanthohoplites aschiltaensis* had to be interpreted by a direct examination and restudy of the two Anthula types mentioned above. The difficulty of my quest in this direction was presented in the last paragraph of my 1949 paper, as below:

> "I have pointed out elsewhere in this paper that Anthula (1899, Pl. 10, fig. 3b) represented the suture of *Acanthohoplites aschiltaensis* with a subsymmetrically bifid first lateral lobe, not in evidence in other species of this genus. In the course of research I tried to locate Anthula's types. This was not an easy task. His described material consisted of two sets: Abich's collection in University of Vienna, at present inaccessible, and Sjögren's collection deposited at University of Upsala. It was in the latter collection, packed and removed to safety during the war, that Dr. H. G. Backlund, Professor in the University of Upsala and Correspondent of the Geological Society of America, assisted by Dr. V. Jaanusson of the same institution, located one of Anthula's (1899, Pl. 10, figs. 2a, 2b) syntypes discussed in this paper, made excellent photographs of the suture, and sent them to me. As I have surmised, the suture is typically "acanthohoplitan," with a symmetrically trifid first lateral lobe and bifid saddles. It is too late to insert an illustration, this will be done in a separate article. I take this opportunity to express my deepest gratitude to Professor Backlund and Dr. Jaanusson for their valued cooperation."

Figures 1 and 2 of Plate 2 are the photographs taken by Doctors Backlund and Jaanusson from Anthula's type of 70 mm diameter, preserved at the University of Upsala, and originally illustrated by him as Figures 2a (this figure is also reproduced in this paper as Figure 4 of

Plate 1) and 2b of his Plate 10. As is inferred from these photographs, the first lateral lobe of this type is tripartite, probably with some asymmetry in its inner branch.

The opportunity to obtain a photographic picture of the holotype, illustrated by Anthula as Figures 3a–3b of his Plate 10 (also reproduced in this paper as Figures 1 and 3 of Plate 1), presented itself when Professor Willis P. Popenoe of the University of California at Los Angeles was doing research work at the University of Vienna in July of 1956 and Professor Othmar Kühn of that University very kindly took a photograph of that type of 45 mm diameter and had it sent to me. This interesting photograph (Pl. 1, fig. 2) shows the exact place on the specimen, marked with a horizontal white line, from which the suture was copied by Anthula. The suture is painted either by india ink or black paint, and the first lateral lobe is fairly well outlined notwithstanding a crack on the orad side of the specimen. One glance suffices to show the discrepancy between this lobe as it appears on the specimen and Anthula's drawing. In the former it is asymmetrically trifid, in the latter asymmetrically bifid. A better-preserved outline of the first lateral lobe is on the orad part of the type, which place I have marked with a vertical white line.

The difference in the nature of the first lateral lobe in the Upsala and in the Vienna types is considerable. In the first, it approximates that appearing in the majority of types described as *Acanthohoplites*; in the second, it suggests Anthula's "allusion" on its proximity to the first lateral lobe in *Parahoplites* which had been contested by Sinzow, as I have mentioned above. Do the types of Anthula belong to two different species?

Not necessarily. The apparent solution comes from a fragment described by Sinzow (1907, p. 479, 516, Pl. 5, fig. 3) as one of two types of his variety, *Acanthohoplites aschiltaensis* var. *rotundata* from Mangyshlak, which I have re-illustrated as Figure 5 of Plate 1. Incidentally, this is the same specimen from which Sinzow derived his interpretation of the difference of the first lateral lobe in *Acanthohoplites* and *Parahoplites*, as has been quoted. Observing the suture of this specimen the reader will note the following changes in the symmetry of the first lateral lobe: (1) symmetrically trifid at the apicad end of the fragment and (2) in the next suture; (3) slightly asymmetrical in the third; (4) more asymmetrical in the fourth; (5) nearly symmetrical in the fifth; (6) strongly asymmetrical in the sixth. This deviation from the symmetrically tripartite first lateral lobe occurs in a few other described specimens referable to *Acanthohoplites*.

Regarding *Parahoplites melchioris* Anthula (1899, p. 112, Pl. 8, figs. 4a–5b), the genotype, it should be noted that his description of the first lateral lobe in the generic diagnosis of *Parahoplites* (Anthula, 1899, p. 111–112; Stoyanow, 1949, p. 96, translated): "The first lateral lobe, which is *very stout and broad* (italics mine), usually is appreciably deeper than the external lobe, it is trifid and with *the external branch stronger developed* than the somewhat smaller inner branch," matches better the examples of *Parahoplites* from Mangyshlak illustrated by Sinzow (1907, Pl. 1, figs. 1, 3, 4, 10; Pl. 2, figs. 1, 2, 4, 5, 9, 11; Stoyanow, 1949, Pl. 17, figs. 9, 10), in which the assymmetrically trifid first lateral lobe invariably has the joined middle and inner branches more strongly developed than the more individualized external branch, whereas

PLATE 1.—*ACANTHOHOPLITES* TYPES OF ANTHULA AND SINZOW

FIGURE

1–4.—*Acanthohoplites aschiltaensis* (Anthula)

(1). Holotype by subsequent designation. University of Vienna. As previously figured by Anthula (1899, Pl. 10, fig. 3a). × 1

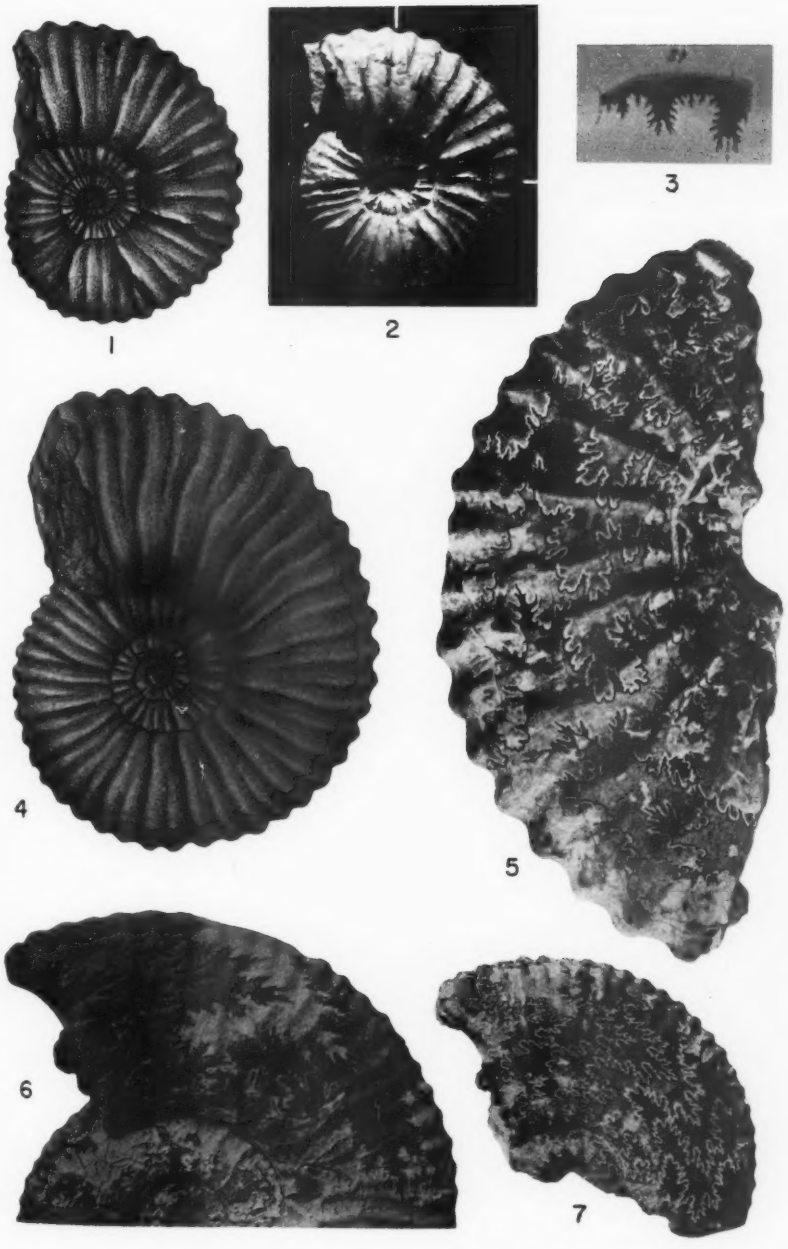
(2). Same. Photograph taken by Professor Kühn. Horizontal white line indicates the suture painted by Anthula. Vertical white line shows the place with a better preserved impression of the first lateral lobe. × 0.91

(3). Same. Suture as originally drawn by Anthula (1899, Pl. 10, fig. 3b)

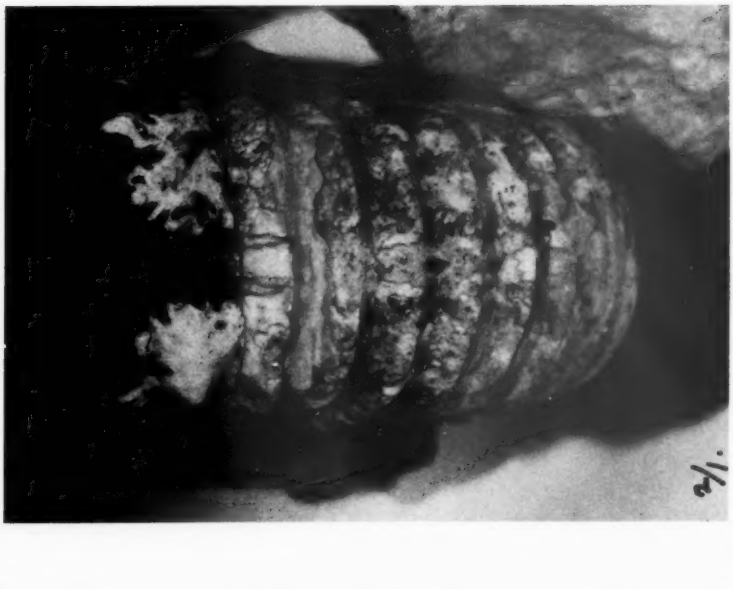
(4). Paratype? As previously figured by Anthula (1899, Pl. 10, fig. 2a). Same type as photographically illustrated in Figures 1 and 2 of Plate 2. × 1

5.—*Acanthohoplites aschiltaensis* var. *rotundata* Sinzow (1907, Pl. 5, fig. 3). To show irregularities in the symmetry of the first lateral lobe in *Acanthohoplites*. From the original photograph by Sinzow. × 1.83

6–7.—Examples of *Acanthohoplites* with characteristically symmetrical trifid first lateral lobe. From original photographs by Sinzow (1907, Pl. 8, fig. 1; Pl. 4, fig. 15). Figure 6—× 0.33; Figure 7—× 1.08



ACANTHOHOPLITES TYPE OF ANTHULA AND SINZOW



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1

ACANTHOHOPLITES ASCHILTAENSIS (ANTHULA)

the suture depicted by Anthula, Figure 4c of his Plate 8, shows the first lateral lobe as very narrow and slender, and without any branch of one side larger than on the other; moreover, these lobes are not alike on the right and left flanks, the right one is more asymmetrical. As re-illustrated by Roman (1938, p. 350, Fig. 325), the left first-lateral lobe in *P. melchioris* is nearly symmetrically tripartite, while in the right one the middle and the external branches together are divided into three subequal parts, whereas the inner branch is blunt and undifferentiated. On the other hand, the illustration by Moore *et al.* (1957, pt. L, 4, p. L386, Fig. 1c) shows the same first-lateral lobes of the right and left sides with fine lobules, and without any appreciably individualized branching. Unfortunately, the fine photograph of *P. melchioris* sent to me by Doctor Kühn is that of the larger of Anthula's types, which does not show the suture either on the side illustrated by Anthula, Figure 5a of his Plate 8, or on the opposite side, the photograph of which I have.

The conclusion is that the deviation from the symmetrically tripartite first-lateral lobe of *Acanthohoplites*, thus far observed, does not attain the strong asymmetry of the first lateral lobe in the *Parahoplites* examples described by Sinzow. As to the first-lateral lobe of *P. melchioris* Anthula, *sensu stricto*, the discrepancy between his characterization and the illustration, or in relation to Sinzow's numerous types from Mangyshlak, can be settled only by re-examination of his type with the suture. According to the letter from Professor Backlund of February 10, 1949, this type is not in Sjögren's collection in Upsala; whether it is in Abich's old collection at the University of Vienna, I do not know at the time of this writing.

A few words in connection with the derivation and correct spelling of the generic names of certain Acanthohoplitinae: *Acanthohoplites* Sinzow, *Hypacanthoplites* Spath, and *Paracanthohoplites* Stoyanow. Kilian (1913, p. 345–346), Spath (1923, p. 4), Renngarten (1926, p. 25), and Roman (1938, p. 348) misspelled the name of Sinzow's genus as "*Acanthoplies*," and Spath on page 64 of his 1923 publication introduced a new genus *Hypacanthoplites*, apparently based on that misspelled name. However, in a later publication, referring to Sinzow's genus, Spath (1939, p. 238) stated: "Often contracted to *Acanthoplites*, but the original spelling must be adhered to." Presuming that Spath had derived his generic name from a misspelled word, I (Stoyanow, 1949, p. 32–33, 96) was using "*Hypacanthohoplites*" in my paper on the Lower Cretaceous stratigraphy of southeastern Arizona.

In a note: "Illegitimate 'corrections' of orthography of generic names," C. W. Wright (1950, p. 506) said that my reference to "*Acanthoplites*" as incorrect is justified solely on the basis of Rule 19, and that "*Acanthoplites*" is quite correct etymologically, but it is my opinion that his etymology was not based on a solid foundation. Wright quite arbitrarily derived the "stem" *acantho* from *acanthos* which is a very specialized noun denoting plant-imitation in the Corinthian capitals and which seldom had been used otherwise. Names like *Acanthoceras* (horn with thorns) and *Acanthohoplites* (*Hoplites* with thorns) have their first part derived not from *acanthos*, as Wright believes, but from *acantha* (Yakovlev, 1932, p. 231) which word primarily means a thorn or a spine, a feature connected with the morphology of those ammonites, whereas *acantho* is not a "stem" but merely a combination part taken from the same *acantha* (see Oxford English Dictionary, 1933, vol. 1, p. 49), and of course retaining "o" before the consonants. Had Sinzow in view Corinthian columns or thorny plants of ancient Greece when he was coining a name for his genus? Not at all, he wanted to bring out in the new name his *Knoten* and "*Hoecker*" (tubercles and thorns) by which *Acanthohoplites* is easily distinguished from *Parahoplites*. Wright maintains that "in the formation of compound names from Greek the parts should be compounded as nearly as possible in the way the Greek would have compounded them" [sic]. As a matter of fact, however, such terms, as used in natural science, come to us *via* New Latin. The names of Greek origin in zoology, and of course in paleontology, must follow Latin transliteration rather than the Greek grammar (Van Bellen, 1950, p. 504).

PLATE 2.—*ACANTHOHOPLITES ASCHILTAENSIS* (ANTHULA)

FIGURE

1–2.—*Acanthohoplites aschiltaensis* (Anthula). Paratype? University of Upsala. Photographs taken by Drs. Backlund and Jaanusson from the type originally figured by Anthula (1899, Pl. 10, fig. 2a; also see Pl. 1, fig. 4).

(1). Lateral view to show the symmetry of the trifid first lateral lobe. × 2.

(2). Ventral view showing the external lobe and symmetrically bifid lateral saddles. × 2

Consequently, having read in the same note of Wright that . . . "*Acanthohoplites* is etymologically wrong but nomenclatorially valid . . ." and, "Stoyanow cannot 'correct' Spath's genus name to *Hypacanthohoplites* . . ." and, "Stoyanow's new genera *Paracanthohoplites* etc., are etymologically wrong but nomenclatorially valid. . . ." I was considerably surprised to note in the *Treatise on Invertebrate Paleontology* (Moore, *et al.*, 1957, p. L387), of which Wright was one of co-authors, the alteration of the last name into "*Paracanthoplites*." If, according to Wright, *Acanthohoplites* has superiority over "*Acanthoplites*" in spite of etymology, so has then *Paracanthohoplites* over "*Paracanthoplites*," even if Wright's etymology were correct, and certainly *vice versa*. Fortunately, Rule 19 supersedes the etymological speculations, and the correct names of discussed genera are: *Acanthohoplites*, *Hypacanthoplites*, and *Paracanthohoplites*. The latter genus is in no way connected with *Hypacanthoplites* Spath, morphologically, stratigraphically, or, in the name, which consists of the prefix *Para* (beside) and Sinzow's legitimate generic name *Acanthohoplites*. The compound name is rather long but etymologically correct and to the point.

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GULF OF ALASKA TROUGH PARALLELS ALEUTIAN TRENCH

BY WILLIAM M. GIBSON

Deep-sea sounding lines run across the Gulf of Alaska by Coast and Geodetic Survey ships in recent years indicated possible existence of an extensive shallow trough coursing southwestward from the head of the Gulf of Alaska. Profiles obtained the last 2 years confirm the continuity of this feature which parallels the Aleutian Trench at a distance of about 60 miles for a distance of 250 miles. A 25-mile gap in sounding lines south of the northerly chain of seamounts prevents tracing the southern limits of this trough. There are indications, however, that it extends 30 miles farther south before curving into the Aleutian Trench, north of the middle chain of seamounts and southeast of Kodiak Island, for a total distance of 400 miles.

This scar on the ocean floor is called a trough because its depth is small relative to its width. It is 500 to 700 feet deep and generally 2 to 3 miles wide over most of its distance. It might be called a deep-sea channel or fault trench, but its cross section favors use of the term "trough".

The trough begins near the bottom of the continental slope south of Yakutat Bay in an area covered by old sounding lines which were without benefit of graphic recorders. It is apparent that the depressed area has considerable breadth here and narrows to 6 miles in a distance of 100 miles. The rest of the trough is narrow and has a steeper west slope. Its origin adjacent to the fronts of former glaciers, its broad width off the continental slope, and the simple graded structure on its southwestward trend might indicate that it is a sea channel. It slopes down to the west about 16 feet per mile. Huge volumes of interglacial melt-water might have flowed along the channel. The formation of sea channels off the West Coast, at the entrance to the Bay of Bengal in the Indian Ocean, off the Hudson submarine canyon, and in other North Atlantic and North Pacific areas have been credited to high-density flow created by suspended sediment, high salinity, or low temperature (Ericson *et al.*, 1951; Buffington, 1952; Dietz, 1953; Menard, 1955). However, it is not likely that a trough of this cross section and negligible gradient was formed by turbidity currents extending more than 300 miles from shore.

There is support for a suggestion of tectonic origin of this trough. It borders the major geosyncline of the Aleutian Trench, and its axis follows a reverse curve in conformity with the trend of the tectonic elements mapped on shore by the U. S. Geological Survey. There are indications still to be verified that a swell branches off into the north from near the upper limits of the trough, where it narrows to a 6-mile width.

This newly discovered trough may be a tectonic element marking the southeast edge of the geosyncline of the Aleutian Trench. The inferred parallel extension of the Chugach-St. Elias fault may mark the northern edge. That fault constitutes a northerly and easterly boundary of the Gulf of Alaska Tertiary Province and has been mapped from Lituya Bay to the Copper River Delta, and inferred to curve to the southwest along the continental shelf beneath the sea to cross Kodiak Island immediately north of its southern coast line (Payne, 1955).

Although there are well-known instances of subsidence at Yakutat Bay following a strong earthquake in recent years and at Pamplona Searidge (Personal communication, G. F. Jordan, 1957) these may be associated with subsidence near the axis (extended) of the Aleutian Trench. This subsidence and the uplift of the Gulf of Alaska Tertiary Province along the margins of the geosyncline would fit a pattern of deformation consisting of downwarping in the zone of weakness and high-angle thrust faulting along the margins attending the lateral shortening and folding of the geosyncline prism (Daly, 1951). The flanking uplifted Tertiary Province is composed of a maximum of 25,000 feet of sediment. That the deformation is Recent is attested by unglaciated marine terraces 700 feet above sea level in the vicinity of Lituya Bay (Don J. Miller, personal communication, 1957).

The existence of similar troughs trending southwestward from the northeastern part of the Gulf of Alaska is indicated but must await confirmation of additional graphically recorded sounding lines and other investigations.

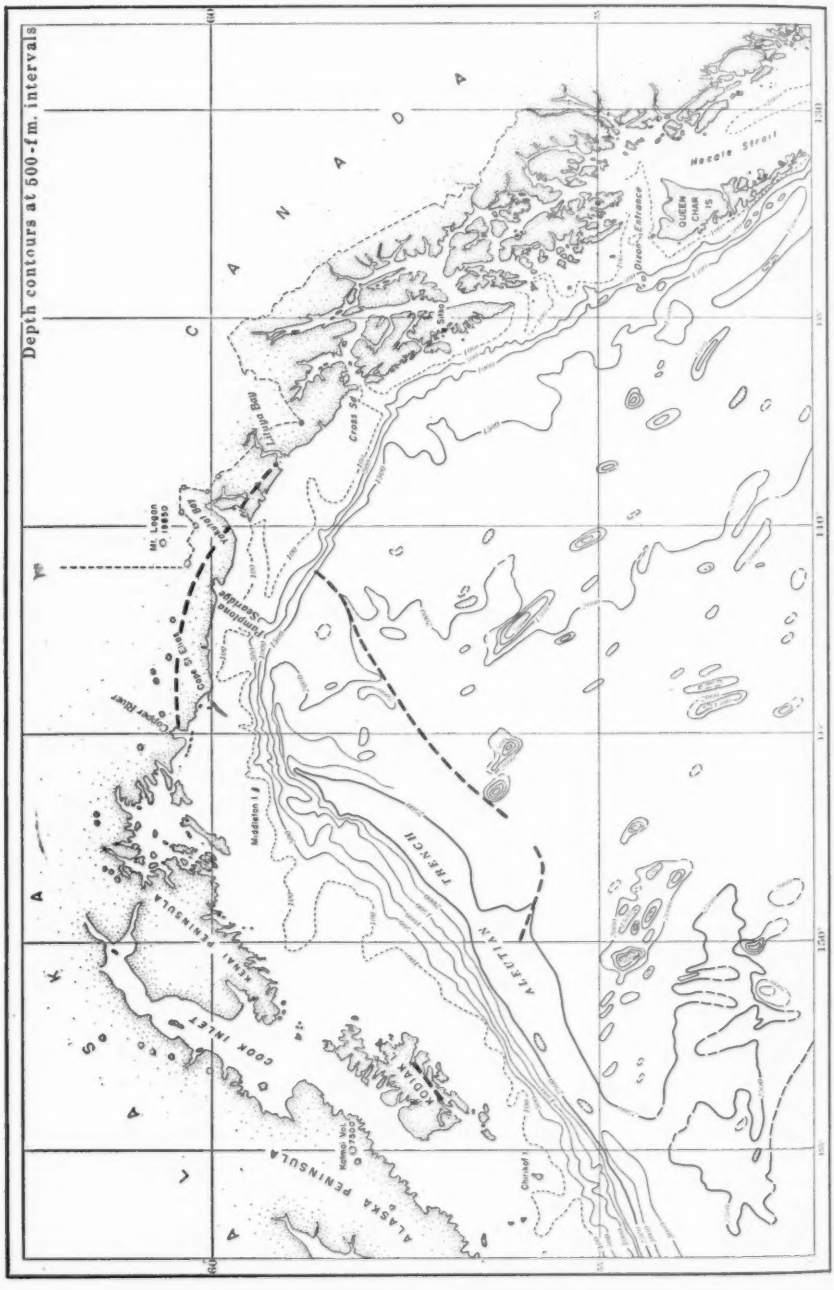

FIGURE 1.—BATHYMETRIC CHART SHOWING AXIS OF EXTENSIVE TROUGH RECENTLY DISCOVERED IN GULF OF ALASKA BY COAST AND GEODETIC SURVEY SHIPS

Also delineated is the inshore change in the 600 fathom bathymetric contour by the U. S. Geological Survey.

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Figure 1.—Bathymetric Chart Showing Axis of Extensive Trough Recently Discovered in Gulf of Alaska by Coast and Geodetic Survey Ships
Also delineated is the highest Quaternary strandline so far identified in the area. (Compiled by the U. S. Geological Survey)

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MILITARY USE OF GEOLOGISTS IN WORLD WAR II

BY R. L. NACE

A recent paper by Kaye (1957) and short notes by Snyder (1957) and Simon (1957) summarize the use of geologists in planning the Normandy invasion and in subsequent military operations on the continent. The authors indicate that geologists served chiefly in staff studies at the Theater Headquarters, as "officer personnel in a water-supply unit," and in photographic interpretation. Briefly, the paper and short notes show that geologists gave important service but that relatively few were used by people who were responsible for employing all suitable talents to accomplish military objectives. But that is not the whole story, and to leave it there is to leave an unfair impression of military ineptitude in the use of geologists. The military importance of geology and related sciences was and is widely appreciated by high military officers in the United States. Appreciation was less acute at successively lower echelons, as might be expected.

Whitmore (1951, p. 2) recognized three general levels of development of geologic information for military purposes: those of intelligence, operational planning, and field consulting. The first is a "high-command" function, the second a theater-command function, and the third a tactical and operational function. The military use of geologists or other professionals can be evaluated properly only on the basis of those three levels of information development.

In the European Theater, little direct operational use was made of geologic talents. At the Theater (operational-planning) level, somewhat more use was made, as Kaye showed. The "high-command" level in the United States appreciated keenly the military importance of geology. For intelligence purposes, the Chief of Engineers, U. S. Army, used the largest concentration of geologic experience in the country (Whitmore, 1954, p. 213), namely, the Geological Survey, and specifically its Military Geology Branch. The intelligence studies made by that unit were tremendously important in high-level military planning and benefitted all lower levels. Some of the intelligence reports themselves reached lower echelons, where the information they contained was used directly.

Hunt (1950) summarized very capably the work of the Military Geology Unit in the United States Geological Survey, but he did not give a chronological account of its development. Hence, it is not widely known that, as early as 1941, the Geological Survey had a definite program of co-operation with the armed services and other government agencies concerned with National Defense. Terrain-intelligence studies and special geologic reports were prepared, and a considerable variety and volume of mineral- and water-resources studies were made. The volume of work increased rapidly, and the Military Geology Unit was created early in 1942. Within a few years the volume of work led to creation of a full-fledged Branch of Military Geology in the Geological Survey. The Branch is still vigorous.

One of the most important military services of geologists was to prepare hundreds of geologic intelligence maps and reports of many kinds, based largely on published information. These staff studies were far superior to any that could have been turned out by one or a few men in a war theater, because the staffs had access to superb libraries in the United States, and their members had a wide variety of skills and a wealth of experiences.

Continued use of geologists and other earth scientists by the military services since World War II also should be noted. The Corps of Engineers regularly uses Terrain Teams composed of commissioned and enlisted earth scientists. Furthermore, geologists, soil scientists, and botanists in uniform have been assigned to work, under the supervision of the Military Geology Branch, in the European and Far Eastern Theaters (Frank C. Whitmore, Jr., personal communication, December 10, 1957). Dott (1957, p. 8) has described good use of earth scientists by the Air Force. Evidently the military services have made considerable progress in recent years in their use of earth scientists. The same is true of the Central Intelligence Agency.

The geologic profession is valued either for military or civilian purposes about in proportion to its practical performance. Proportionately few geologists can or do give prompt concrete answers to specific questions. Too often their answers are inconclusive or profusely

qualified. Too often they demand time for traditional and inappropriate types of studies on which reports become available only after their value for military or engineering purposes has vanished.

The relatively limited use of geologists, both in military and civilian engineering problems, is largely the fault of geologists themselves. Many geologists are extremely reluctant to break with certain traditions, either in response to military alarums and excursions or for workaday purposes. Only in recent years, for example, have they begun to get widened recognition by the engineering profession. Military geology is largely engineering geology, and geologists as a group failed for decades to sell themselves in that field or even to take it very seriously. Hunt's (1950) paper on military geology merits thoughtful reading by all who are concerned with this problem, and the "challenge" described by Dott (1957) deserves serious attention.

Incidentally, the Military Geology Branch consistently takes appropriate opportunities to gain wider recognition of the military uses of earth sciences and also to get better use of earth scientists in uniform.

The references by Simon (1957) and Kaye (1957, p. 53) to use of "a water-supply unit" in the European Theater convey an inaccurate impression that operational use of geologists was practically nil. Kaye (p. 50) refers to this unit as one that operated with the Seventh Army in southeastern and eastern France. That company was commanded by R. C. Newcomb, a Geological Survey geologist. In addition, the 487th Water-Supply Battalion, consisting of a Headquarters and Service Company and three operating companies, served in the northern sector of the theater. Although only one company was commanded by a geologist, the writer, his geologic knowledge was available to and used by other companies, by engineer construction units, and by civilian water-supply engineers in France and Belgium. The writer knows of at least one combat engineer unit commanded by a geologist. Frank Foley, a Survey geologist, was executive officer of a construction battalion in Africa and Italy. W. C. Rasmussen, who won the Bronze Star at Anzio, commanded a water-supply company that drilled more than 300 wells for the Fifth Army in Italy. (A. N. Sayre, written communication, January 21, 1957). Another Survey geologist, R. C. Cady, was killed in North Africa while doing military geologic and water-supply work.

Further search undoubtedly would disclose substantial professional contributions to the war effort by geologists in the European theater, as well as in other theaters. Therefore, to round out the picture, it seems pertinent to add a few highlights about the Pacific theater. There, several officer geologists were attached to General Headquarters, mainly for terrain-intelligence work. Two teams of Geological Survey geologists, one with GHQ, Southwest Pacific Area, and one with the Tenth Army, made terrain studies for beach-head areas on Leyte, Okinawa, and other islands. Several of these men served later as geologists with troops in combat zones. At least two geologists commanded construction battalions in that theater. Two Survey geologists, D. J. Cederstrom and H. T. Stearns, were used widely by the Navy on water-supply and engineering problems at many Pacific bases. Others, among them A. N. Sayre, in addition to their work in the Military Geology Unit, went on extensive overseas missions that had profound strategic and tactical significance.

In evaluating the military use of earth scientists, the fluid nature of global war should be remembered. High-level use of the best brains and widest experience affects strategy and broad tactics. Low-level use, of less-experienced men, affects field tactics. The latter influences a skirmish or battle. The former may affect the outcome of a war.

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